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**Geologic Setting, Geochronologic Relationships, and Lithogeochemistry
of the Pangui Porphyry Copper District, Southeast Ecuador.**

by

Joshua McDermott Coder



A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment
of the requirements for the degree of Master of Science.

Department of Earth and Atmospheric Sciences

Edmonton, Alberta

Fall, 2001

University of Alberta

Faculty of Graduate Studies and Research

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled Geologic Setting, Geochronologic Relationships, and Lithogeochemistry of the Pangui Porphyry Copper District, Southeast Ecuador. Submitted by Joshua McDermott Coder in partial fulfillment of the requirements for the degree of Master of Science.

Abstract

The recently discovered Pangui porphyry copper district located in the subandean zone of southeast Ecuador is hosted within the calc-alkaline Zamora Batholith, which is part of a larger volcanic arc that extended from Colombia to Chile. Lithogeochemical data show that the Zamora Batholith, porphyry phases, and Oyotún Fm. are broadly cogenetic. The porphyry copper deposits within the Colombia-Ecuador segment of the volcanic arc were emplaced between 170 and 150 Ma.

10 porphyry copper prospects and 1 copper skarn prospect have been identified within the Pangui District. The porphyry prospects contain high hypogene copper contents and are the result of multiple hydrothermal events. Grade and tonnage estimates are available for the San Carlos and Panantza prospects (230 Mt at 0.85% Cu and 148 Mt at 0.82% Cu respectively). Radiometric age-dating suggests that the mineralization occurred over a period of approximately 5 m.y. during the Late Jurassic.

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TABLE OF CONTENTS

ABSTRACT	iii
ACKNOWLEDGEMENTS	iv
TABLE OF CONTENTS	v
LIST OF TABLES	ix
LIST OF FIGURES	x
1. INTRODUCTION	1
i) Location and previous work	1
ii) Mapping program	3
iii) Lithogeochemistry	3
iv) Radiometric age-dating	3
v) Petrographic work	3
2. UPPER PALEOZOIC AND MESOZOIC TECTONIC HISTORY OF	
NORTHWESTERN SOUTH AMERICA	4
Permian–Triassic rift	4
i) Colombia	4
ii) Perú	6
Early Jurassic–Early Cretaceous Vicusian Orogeny	7
Terrane Accretion along the northwest margin of South America ...	11
i) Late Jurassic–Early Cretaceous	12
ii) Late Cretaceous–Paleocene	15

3. GEOLOGIC HISTORY OF THE CORDILLERA DEL CONDOR	17
Precambrian	17
Paleozoic	20
i) Late Silurian-Mississippian	20
ii) Pennsylvanian	21
Permian–Mesozoic	21
i) Permian–Early Triassic	22
ii) Late Triassic–Early Jurassic	22
iii) Early–Late Middle Jurassic	24
iv) Late Jurassic–Early Cretaceous	26
v) Late Early Cretaceous–Tertiary	26
4. VICUSIAN METALLOGENY	28
Introduction	28
Plate movements and migration of mineralization	28
Jurassic Cretaceous mineralization	30
i) Alteration	35
Cretaceous mineralization	36
i) The Nambija District	37
5. DESCRIPTION OF THE SAN CARLOS AND PANANTZA PROSPECTS	39
Introduction	39
Prospect Geology	40
i) Oyotún Fm.	40
ii) Zamora Batholith	42
Porphyry dike relationships	44

i) Early porphyry dikes	44
ii) Late porphyry dikes	47
iii) Post mineral dikes	47
Hypogene Alteration and Mineralization Characteristics	52
i) Panantza	52
ii) San Carlos	57
iii) Supergene Oxidation and Enrichment	64
6. WHOLE-ROCK LITHOGEOCHEMISTRY OF IGNEOUS ROCKS	
FROM THE PANGUI DISTRICT	66
i) Major and trace element methodology	66
ii) Major elements	66
iii) Oxide-Oxide plots	72
iv) Trace elements	72
7. RADIOMETRIC AGE-DATING OF THE SAN CARLOS, PANANTZA,	
AND SUTSU PROSPECTS	77
Introduction	77
Sample Selection	77
i) $^{40}\text{Ar}/^{39}\text{Ar}$	77
ii) K-Ar	77
iii) Re-Os	78
Analytical Methods	78
i) Principles of the Re-Os technique with respect to molybdenite ...	78
ii) Re-Os methodology	79

ii) $^{40}\text{Ar}/^{39}\text{Ar}$ methodology	80
Geochronologic Results	83
i) $^{40}\text{Ar}/^{39}\text{Ar}$ age results	83
ii) Re-Os age results	85
iii) $^{40}\text{K}/^{40}\text{Ar}$ age results	85
8. DISCUSSION	87
i) Tectonic setting	87
ii) Zamora Batholith	89
iii) Geochemistry of the Jurassic igneous suite	90
iv) Jurassic tectonics and porphyry copper formation	96
v) Metallogeny	99
vi) Pangui district porphyry model	101
vii) Late dacite porphyry	103
iv) Supergene enrichment	104
9. CONCLUSIONS	106

LIST OF TABLES

Table

2.1.	Major tectonic events in northwestern South America from the Early Triassic through the Paleocene	9
4.1.	List of prospects	34
5.1	Paragenesis of the Panantza prospect	55
5.2	Paragenesis of the San Carlos prospect	60
6.1.	Major and trace element data from the San Carlos prospect	67
6.2.	Major and trace element data from the Panantza prospect	68
7.1.	$^{40}\text{Ar}/^{39}\text{Ar}$ data	82
7.2.	Re-Os dates	86
7.3.	K-Ar dates	86

LIST OF FIGURES

Figure

1.1.	Map showing the location of the Cordillera del Condor, Cordillera de Cutucu, and concession blocks	2
2.1.	Regional sketch map of western Pangea for the Permian through Triassic showing the location of rift centers and aulacogens	5
2.2.	Regional sketch map of western Gondwana and Laurasia for the Late Triassic through Early Jurassic showing the post rift marine transgression and the initiation of subduction	8
2.3.	Regional sketch map of western Gondwana and North America for the Early through Middle Jurassic showing the location of the volcanic arc	10
2.4.	Regional sketch map of western South and North America for the Early through Late Cretaceous showing the accretion of continental and oceanic terranes onto the western margin of South America	13
3.1.	Geologic interpretation map of the Cordillera del Condor, Cordillera de Cutucu, and north-central and northwestern Perú	18
3.2.	Correlation diagram of the stratigraphy for the Cordillera del Condor and Cordillera de Cutucu between Ecuador and Perú	19
4.1.	Map showing the locations and spatial associations of the Jurassic through Neogene volcanic arcs along the western margin of South America	29
4.2.	Map showing the locations of Jurassic through Miocene prospects and	

districts in Colombia, Ecuador, and northern Perú	31
4.3. Map showing the location of the individual prospects within the Pangui District, Southeast Ecuador	33
5.1. Photographs of hand samples and thin sections of the Oyotún Fm.	41
5.2. Photographs of hand sample and thin section of the Pink Zamora Batholith	43
5.3. Photograph of the contact between the Zamora Batholith and Oyotún Fm.	44
5.4. Photographs of hand samples and thin sections of the leucocratic granite at the San Carlos prospect	45
5.5. Photographs of hand samples and thin sections of the early porphyry	46
5.6. Photographs of hand samples and a thin section of the late porphyry	48
5.7. Photographs of a weathered hand sample of the post-mineral porphyry dike	49
5.8. Photographs of hand samples and thin sections of the post-mineral microporphyritic andesite	50
5.9. Photographs of hand samples and thin sections of the post-mineral rhyolite	51
5.10. Sketch interpretation map of alteration at the Panantza prospect	53
5.11. Sketch interpretation map of alteration at the San Carlos prospect	54

5.12.	Oyotún Fm. andesite hosting veins of phyllic alteration	58
5.13.	Photograph of hand sample and thin section of a Pb-Zn-Cu vein from the Panantza prospect	59
5.14.	Photograph of a hand sample of early porphyry showing K-spar overprinting at the San Carlos prospect	61
5.15.	Photograph of a hand sample and thin section of coarse-grained muscovite and quartz alteration from the San Carlos prospect	63
6.1a	Ternary diagram after Irvine and Baragar (1971) showing a calc-alkaline trend	70
6.1b	TAS diagram showing alkaline, midalkaline, and subalkaline trends ...	70
6.2a	TiO ₂ -Al ₂ O ₃ diagram after Muller and Groves (1993) showing and arc-related trend	71
6.2b	QAP Diagram after Streckeisen (1976)	71
6.3.	K ₂ O vs SiO ₂ diagram after LeMaitre (1989) showing medium to high K affinities of the sample suite	72
6.4.	Oxide versus SiO ₂ plots	73
6.5a	Primitive mantle normalized plots	74
6.5b	Primitive mantle normalized plots	75
6.6	C1 normalized rare earth element plots	76
7.1.	Laser ⁴⁰ Ar/ ³⁹ Ar inverse isochrons diagrams and step plots	84
8.1.	Tectonic discrimination diagrams	88
8.2a	Plot of Eu-anomaly versus SiO ₂	93
8.2b	Plot of La/Sm versus La/Yb	93

8.3.	Plot of Th versus $1/Yb$		95
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1. Introduction

The central objective of this work has been to provide a baseline characterization of the Pangui porphyry copper district, SE Ecuador. In order to achieve this objective a program of geologic mapping supported by lithogeochemistry, geochronology, transmitted and reflected light petrographic work, and an in-depth review of the literature was carried out.

Location and previous work

The study area is located in southeastern Ecuador along the border with Perú in the provinces of Morona-Santiago and Zamora-Chinchipe within the Cordillera del Condor, just to the south of the Cordillera de Cutucu (Fig. 1.1). This region is a densely vegetated tropical mountainous terrain of up to 2000m elevation.

Gendall et al. (2000) published a description of the exploration history of the district from which the following synopsis has been taken. In 1994 a GENCOR subsidiary discovered the Pangui porphyry copper district while prospecting for gold. The Pangui district covers an area of 1640 km² and hosts 10 known porphyry copper prospects and one copper skarn in a 70-km long belt covering the northern and central parts of the Cordillera del Condor. Drilling on a few of the prospects has confirmed the existence of porphyry copper type mineralization. By 1999 Billiton formed a joint venture for exploration within the district with Corriente Resources and Lowell Mineral Exploration. Since 1999 the joint venture partners have drilled a number of the known porphyry prospects but no new porphyry prospects have been discovered.

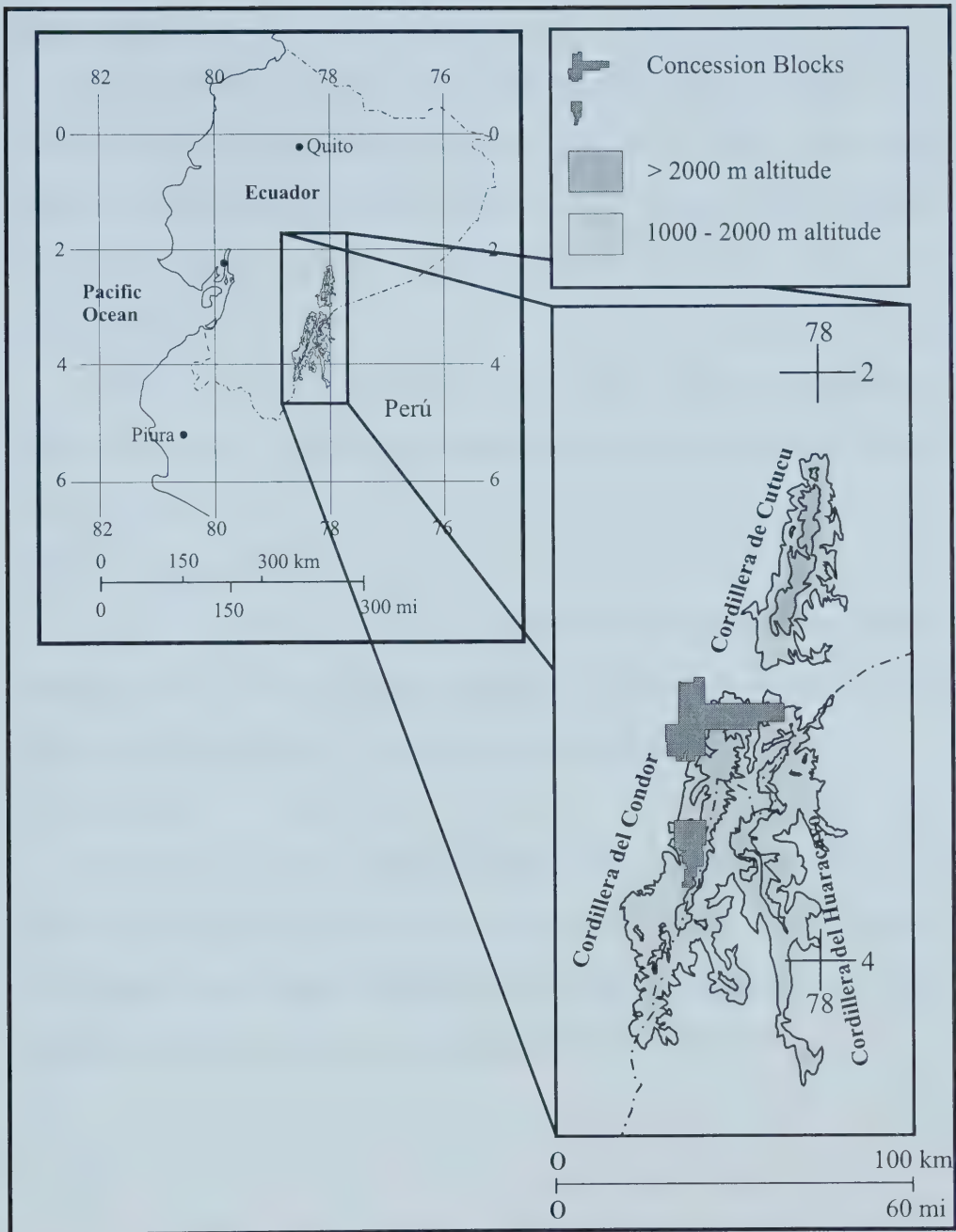


Fig. 1.1. Map of Ecuador and Perú showing the location of the Cordillera del Condor, Cordillera de Cutucu, and concession blocks. The study area is located within the northern concession block.

Mapping program

Fieldwork was carried out by the author over two field seasons (two to three months each) during 1998 and 1999, using field vehicles, helicopters, mules, support personnel, and camps provided by Billiton Ecuador, B.V. Mapping was conducted at a scale of 1:10,000 using topographic maps based on the UTM grid 17 Sur PASD56.

Lithogeochemistry

A suite of variably altered rocks was analyzed for major and trace elements in order to ascertain the tectonic setting and relationships between the various units present in the study area.

Radiometric age-dating

Samples of alteration minerals, molybdenite, and hornblende from igneous rocks have been dated to provide information on the duration of hydrothermal activity and the relationship of various porphyry units to alteration and mineralization events.

Petrographic work

Transmitted and reflected light petrographic work was carried out on a suite of thin sections in order to determine the various relationships between primary, alteration, and secondary mineral phases. Identification of intrusive and volcanic rock types based the IUGS classification scheme was also undertaken where possible.

2. Upper Paleozoic and Mesozoic Tectonic History of Northwestern South America

Permian-Triassic rift

The northern Andes between 12°N and approximately 8°S saw an intracontinental rift develop during the Middle (?) Permian to Late (?) Triassic. This event was related to the breakup of Pangea, which affected present-day Colombia, Ecuador, and Perú (Irving, 1975; Jaillard et al., 1990; Zamora and Litherland, 1993a; Benavides-Cáceres, 1999).

Two rift centers are apparent: the first was located in northernmost Colombia, and the second was located in the vicinity of Huancabamba in northern Perú (Jaillard et al., 1990). Both rift centers shared the rift arm that strikes NE from the Huancabamba rift center to the rift center in northern Colombia (Jaillard et al., 1990; Fig. 2.1). The rift opened an ocean basin along the western margin of Colombia, Ecuador, and north-westernmost Peru, whereas an aulacogen that struck SE toward the Bolivian and Chilean frontier developed in the rest of Perú (Jaillard et al., 1990).

As the North American plate began to pull away from the South American plate (Fig. 2.1) in a north-northwest direction, different structural styles developed in the Colombian and the Ecuadorian-Peruvian segments of the rift (Jaillard et al., 1990; Mojica and Kammer, 1995). Strongly oblique extension of the rift relative to Colombia formed a basin and range structure, whereas along the Ecuadorian segment the extension was less oblique and core complex structures are preserved (Aleman and Ramos, 2000).

Colombia

In the Colombian rift, the aulacogen strikes northwest into the North American plate, whereas the other two completely rifted arms strike east toward Venezuela and southwest paralleling the western coast of Colombia and Ecuador (Jaillard et al., 1990;

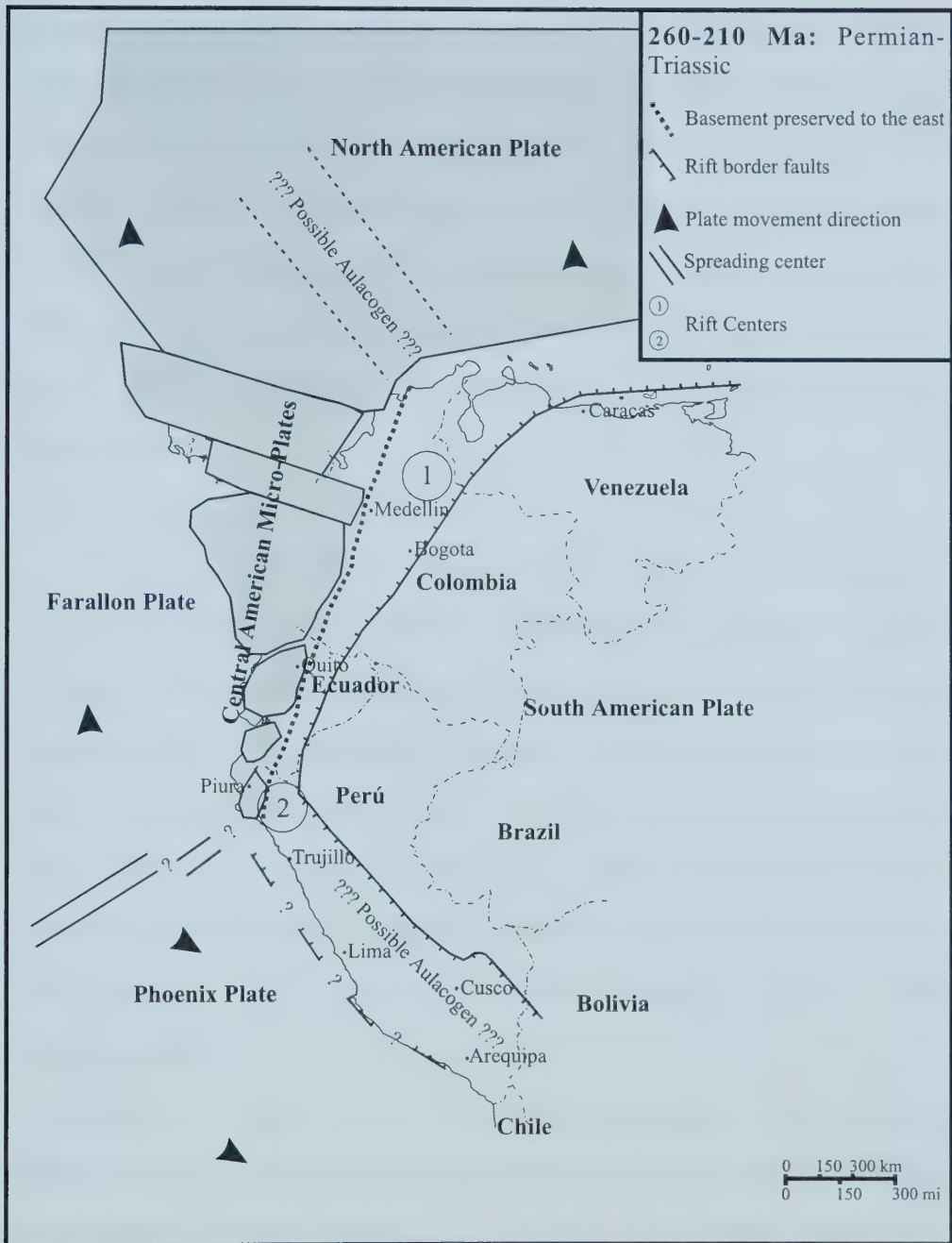


Fig. 2.1. Regional tectonic sketch map of northwest South America showing the location of the two rift centers and probable aulocogens in Perú and North America. Plate movement direction is relative to the South American plate (modified from: Aspdén et al., 1987; Jaillard et al., 1990; Irving, 1975; Zamora and Litherland, 1993a; INGEMMET, 1995 and 2000; Aleman and Ramos, 2000).

Aleman and Ramos, 2000; Fig. 2.1). Along the southwest striking arm in Colombia, scattered deformed diorites, monzonites, and tonalites of Middle to Late Triassic age intrude the western metamorphic belt in the Central Cordillera along a poorly defined NNE strike (Irving, 1975; Toussaint and Restrepo, 1982; Aspden et al., 1987; Jaillard et al., 1990). These deformed intrusions are noted by Aspden et al. (1987) to have a strong NW-SE biotite foliation. In Ecuador, Middle to Late Triassic S-type granitoids crop out along a NE-SW trend that spans the length of the Cordillera Real (Zamora and Litherland, 1993a).

Perú

The second rift center was located in the vicinity of Huancabamba in northwestern Perú. The aulacogen of this center strikes southeast, spanning the length of the country (Jaillard et al., 1990; Fig. 2.1). Two belts of Precambrian and Paleozoic metamorphic rocks are preserved in the aulacogen. The first belt is located along the present day Pacific coast and crops out as far north as Lima, whereas a second belt, further to the east, crops out nearly continuously along a northwest strike from the Bolivian border to the Ecuadorian border: here the belt abruptly changes to a northeast strike in the vicinity of the town of Huancabamba (Zamora and Litherland, 1993a; INGEMMET, 2000).

Synrift sedimentary infill of the aulacogen is comprised by the Mitu Group, a sequence of dark colored sandstones, red-green conglomerates (apparently sourced from the Precambrian Mara  n Complex), and red and purple volcanic rocks that rest discordantly on the Devonian to Mississippian Ambo Group and the Mara  n Complex (Wilson et al., 1967). Wilson et al. (1967) noted that the region underwent extensional

faulting during the deposition of the Mitu Group. They observed abrupt thickness changes in the strata coupled with an increase in a volcanic component toward the upper part of the Group. Carlier et al. (1982) described the volcanic strata in the Mitu Group as outcropping widely between 6°S and 16°S. Two magma types were erupted: an alkaline sequence consisting of welded tuffs, rhyolites, andesites; and a subordinate spilitized tholeiitic basalt sequence (Carlier et al., 1982). These volcanic sequences apparently erupted along large faults during extension in the Eastern Cordillera; a coeval calc-alkaline chain of porphyritic granitoids intruded into the same region (Carlier et al., 1982). These intrusions straddle the Permian-Triassic boundary and young in two directions from central Perú: southward toward the Bolivian border and northward into Ecuador and Colombia (Irving, 1975; Carlier et al., 1982; Toussaint and Restrepo, 1982; Aspden et al., 1987; Jaillard et al., 1990; Zamora and Litherland, 1993a).

Early Jurassic-Early Cretaceous Vicusian Orogeny

During the Early Jurassic, post-rift subsidence caused a marine transgression in the vicinity of Piura in northern Perú that then propagated north into Ecuador and southeast into the failed Peruvian rift arm (Benavides-Cáceres, 1999; Fig. 2.2; Table. 2.1). Subduction commenced during the Latest Triassic - Early Jurassic, and resulted in a calc-alkaline volcanic arc consisting of two segments: Colombia-Ecuador-north-westernmost Perú, and north central Perú-northwest Chile (Fig. 2.3). The volcanic arc was emplaced into continental crust along former rift border faults in Colombia, Ecuador, north-westernmost Perú, southwestern Perú, and northwest Chile (Aspden et al., 1987; Jaillard et al., 1990; Vila et al., 1996; Marschik and Fontbote, 1996).

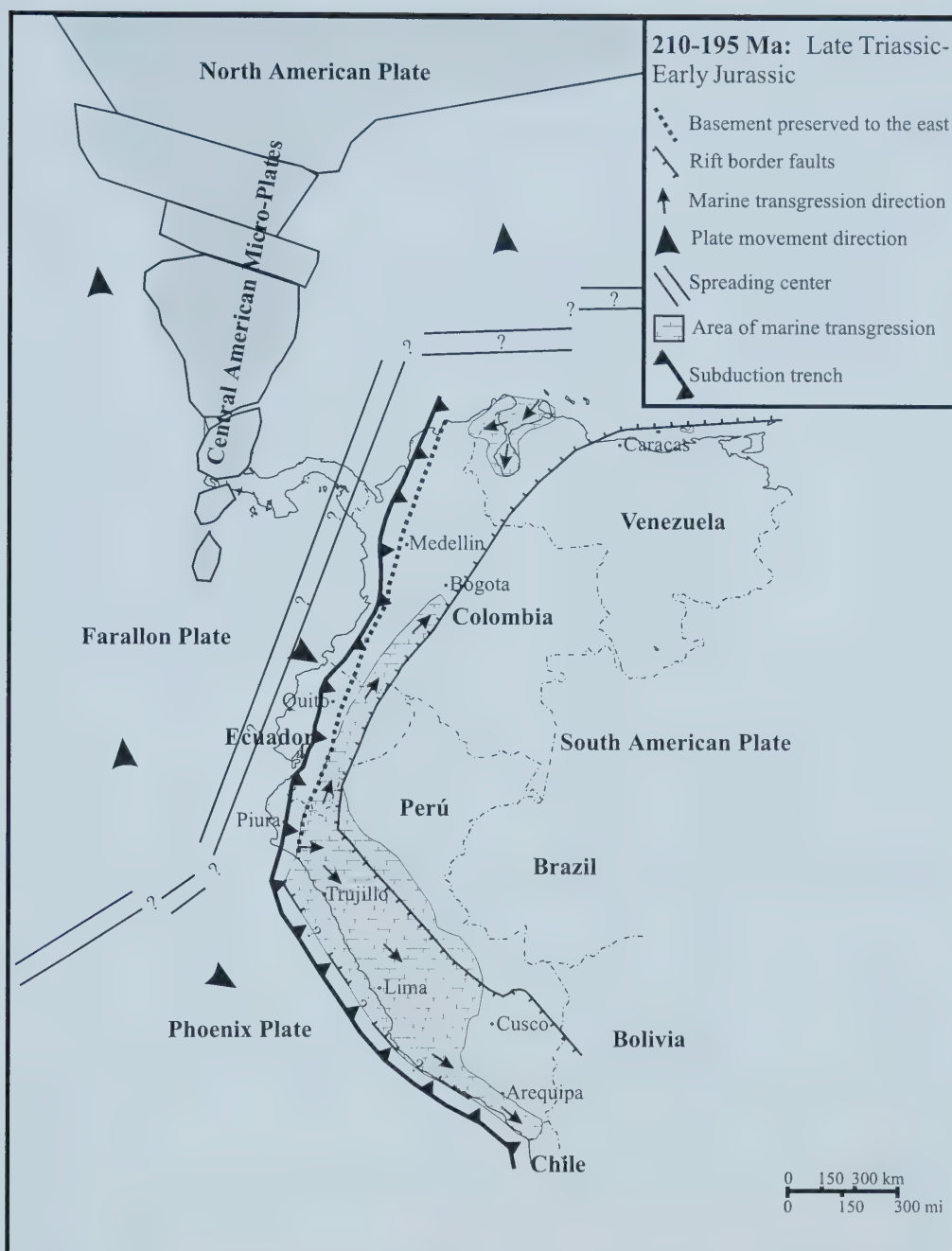


Fig. 2.2. Regional tectonic sketch map of northwest South America showing the location and direction of the marine transgression in Colombia, Ecuador, and Perú. Subduction begins along the western margin of the South American plate (modified from: Aspden et al., 1987; Jaillard et al., 1990; Irving, 1975; Zamora and Litherland, 1993a; INGEMMET, 1995 and 2000; Aleman and Ramos, 2000).

TABLE 2.1. Tectonic framework of the Cordillera del Condor.

Period	Age range (Ma)	Tectonic event	Tectonic and magmatic behavior	Manifestation in Cordillera del Condor	References
Late Cretaceous-Late Paleocene	75-58	Terrane accretion and uplift of subandean zone	Volcanic arc shifts west a second time	Deposition of Tena Fm., uplift and erosion and weak deformation	Toussaint and Restrepo (1982) Aspden et al. (1987) Feininger (1987) Mourier et al. (1988) Zamora and Litherland (1993a) Arculus et al. (1999)
Early Cretaceous-Late Cretaceous	132-75	Terrane accretion and marine transgression (Termination of Viscusian Orogeny)	Uplift and erosion, volcanic arc shifts west	Last burst of magmatic activity deposits the Misahualli unit of the Chapiza Fm., major regional unconformity, Hollin/Gollarisquisga and Napo/Chonta Fms deposited	
Early Jurassic-Early Cretaceous	198-132	Subduction	Subaerial to subaqueous volcanic arc extending from Colombia to Chile, copper mineralization in Colombia, Ecuador, Chile	Intrusion of the Zamora Batholith, deposition of the Chapiza/Sarayaquillo Fms., porphyry copper and skarn deposit formation	Tschopp (1953) Sillitoe et al. (1982) Aspden et al. (1987) Sillitoe (1988) Jailard et al. (1990) Tafur (1991) Zamora and Litherland (1993a) Marschik and Fonbote (1996) Quipesivana (1996) Vila et al. (1996) Benevides-Caceres (1999) Gendall et al. (2000)
Early Jurassic	208-198	Post rift subsidence, subduction commences (Viscusian Orogeny begins)	Initial marine transgression: volcanic arc to west and sedimentary basin to east	Deposition of the Santiago Fm./Pucara Group: volcanic facies to west (Piuntza unit) and carbonate/clastic facies to east	Tschopp (1953) Zamora and Litherland (1993a) Quipesivana (1996) Benevides-Caceres (1999)
Late Triassic-Early Jurassic	230-208	Rifting of North America From South America	Deposition of red beds, tholeiitic basalt, and alkalic volcanics in Perú, intrusion of S-type granites in Ecuador, and scattered deformed intrusives in the Central Cordillera of Colombia	Deposition of Mitu Group in Perú and emplacement of S-type granites in Ecuador	Wilson et al. (1967) Irving (1975) Carlier et al. (1982) Toussaint and Restrepo (1982) Aspden et al. (1987) Jailard et al. (1990) Zamora and Litherland (1993a)
Early Triassic-Late Triassic	250-230	Breakup of Pangea commences	Rift valleys form in Colombia, Ecuador, and Perú.	None known	

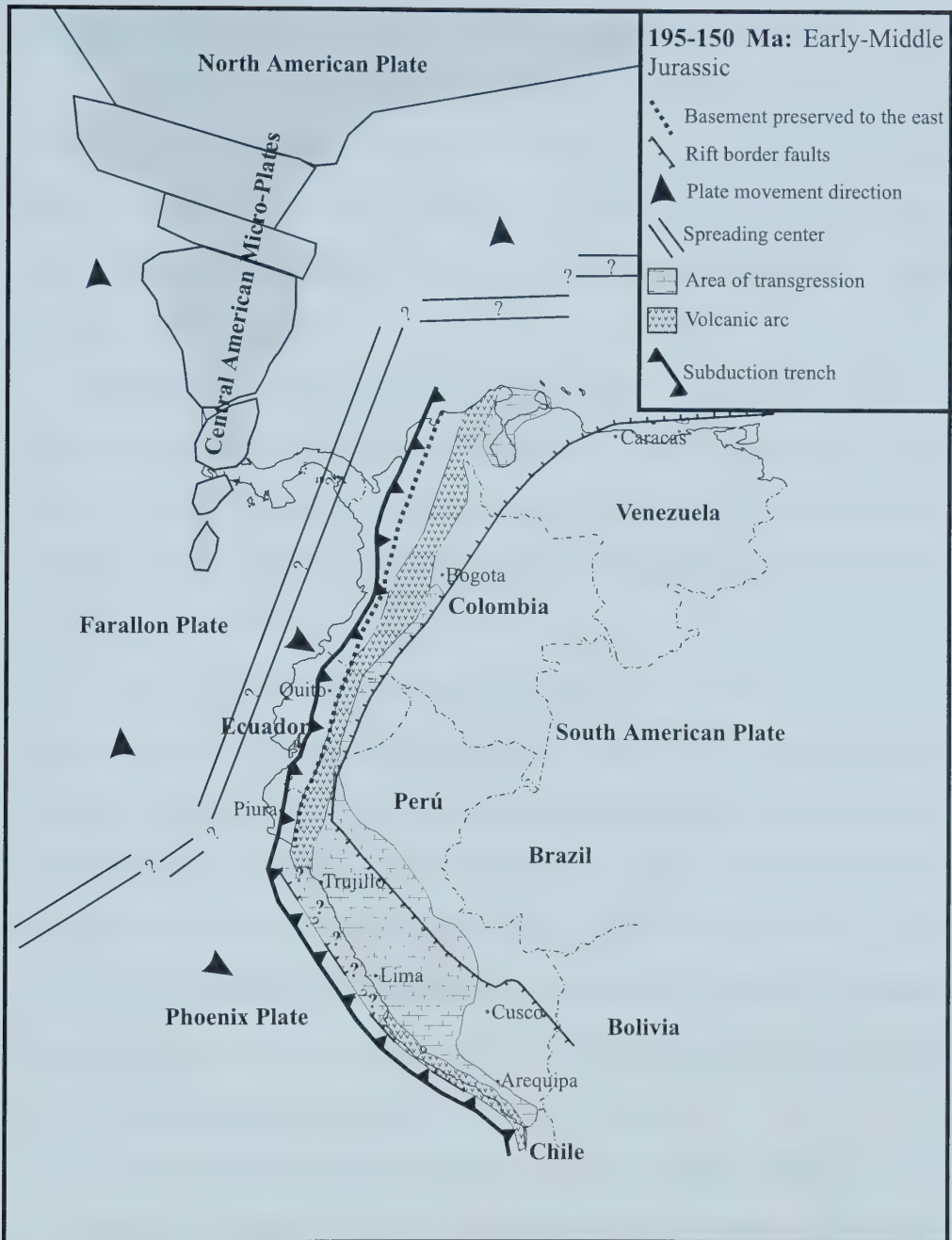


Fig. 2.3. Regional tectonic sketch map of northwest South America showing the location of the Jurassic volcanic arc in Colombia, Ecuador, Perú, and northern Chile (modified from: Aspden et al., 1987; Jaillard et al., 1990; Irving, 1975; Zamora and Litherland, 1993a; INGEMMET, 1995 and 2000; Aleman and Ramos, 2000).

This magmatic event is named the Vicusian Orogeny (Benavides-Cáceres, 1999).

The volcanic arc in Colombia was dominantly subaerial, but exhibits at least a partially subaqueous component further to the south in Ecuador and Perú where it apparently formed a chain of islands in a shallow sea (Jaillard et al., 1990). Further south along the Perú-Chile segment the volcanic arc was subaqueous (Vila et al., 1996; Marschik and Fontbote, 1996).

Magmatism continued through most of the Middle Jurassic, but a hiatus in volcanism locally occurred between 169-158Ma (Tafur, 1991; Quispesivana, 1996; Benavides-Cáceres, 1999). This volcanic hiatus coincides with a period of emplacement of porphyry copper deposits in Colombia, Ecuador, and possibly Perú (Sillitoe et al., 1982; Gemuts et al., 1992; Gendall et al., 2000).

Although uplift of the region probably began in the Early Jurassic, continental sedimentation (while continuing in Colombia) only commenced in Ecuador and Perú at some point during the Middle Jurassic (Zamora and Litherland, 1993a). During the latest Jurassic, subaerial volcanic activity resumed in Perú, Ecuador, and southern Colombia and continued into the Early Cretaceous (Hall and Calle, 1982; Jaillard et al., 1990). Along the Perú-Chile segment of the volcanic arc, volcanism was continuous throughout the entire Jurassic-Early Cretaceous. Accreted terranes are unknown during this period (Vila et al., 1996; Marschik and Fontbote, 1996).

Terrane Accretion along the northwest margin of South America

Much of the following description is summarized from Toussaint and Restrepo (1982), Aspden et al. (1987), Feininger (1987), Mourier et al. (1988), Zamora and Litherland (1993a), and Arculus et al. (1999). Two periods of terrane accretion occurred

along the northwestern margin of South America. Both accretion events, during the Latest Jurassic and/or Early Cretaceous and Late Cretaceous to Paleocene, caused westward jumps in the volcanic arc.

Latest Jurassic-Early Cretaceous

The Amaime, Amotape, Tahuin, and Chaucha terranes were accreted during the Early Cretaceous along the Romeral-Peltetec-Giron-Las Aradas fault system of Colombia, Ecuador, and northwestern Perú (Zamora and Litherland, 1993a; Fig. 2.4). Zamora and Litherland (1993a) interpret the ultramafic and mafic rocks that crop out sporadically along the fault system as an ophiolite sequence and the fault system as a paleo-subduction zone. However, the suture is not as apparent in Perú because it is concealed beneath Cretaceous cover rocks (Zamora and Litherland, 1993a; Feininger, 1987).

The Amotape and Tahuin terranes in Perú and Ecuador are contiguous and will henceforth be designated the Amotape-Tahuin terrane; this terrane includes the Lancones Basin (Feininger, 1987; Arculus et al., 1999). The Amotape-Tahuin terrane is believed to have originated south of its current position straddling the Perú-Ecuador border, and its oldest strata show a clockwise 94° rotation (Mourier et al., 1988). The accretion of this terrane is generally believed to have occurred during the Aptian to Albian, but published ages ranging from the Neocomian to Albian suggest a slightly earlier age of accretion (Feininger, 1987; Mourier et al., 1988; Arculus et al., 1999). Mourier et al. (1988) noted a westward jump of the volcanic arc along the Peruvian-Ecuadorian segment of approximately 150km during the period after the accretion.

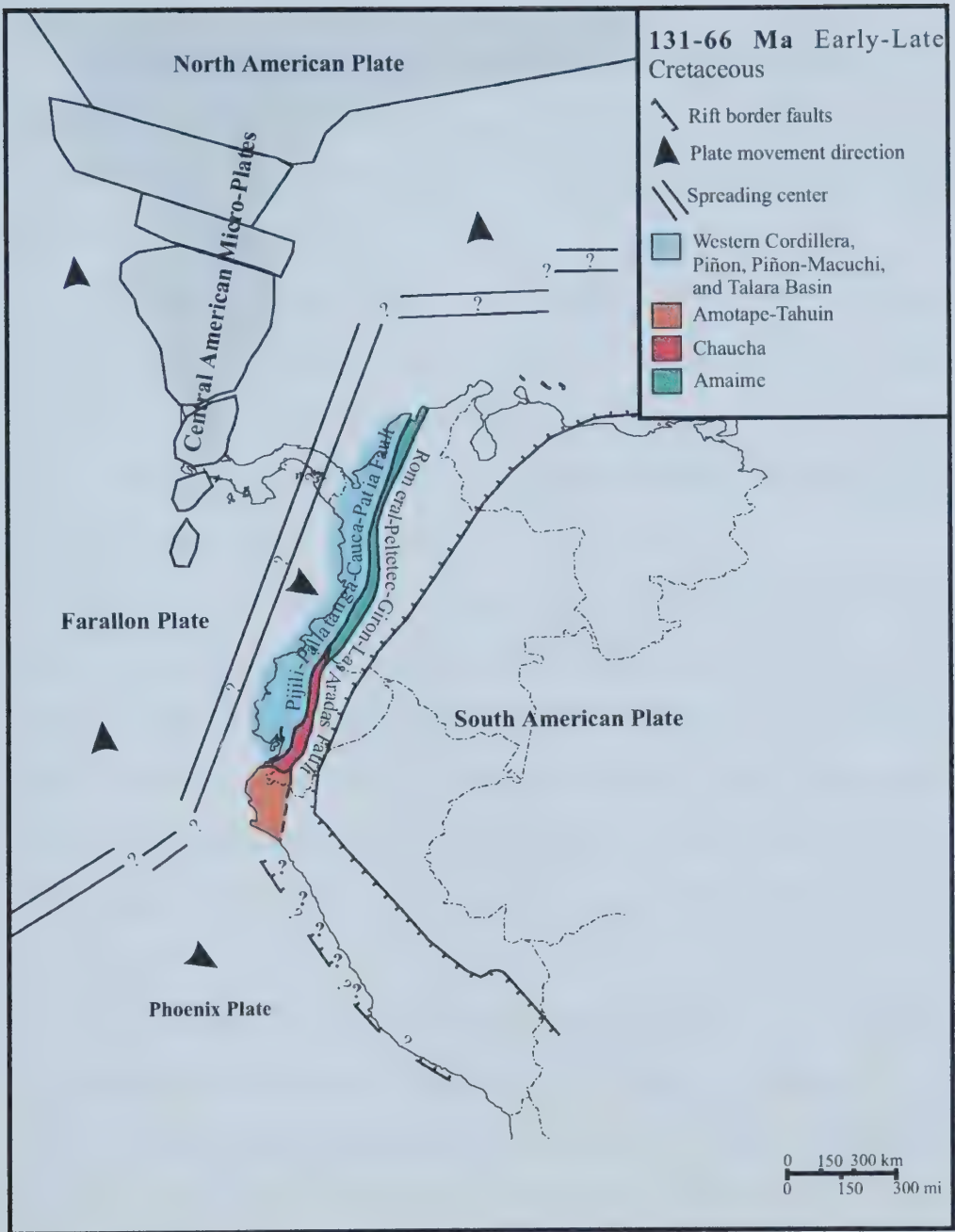


Fig. 2.4. Regional tectonic sketch map of northwest South America showing the location of the Early and Late Cretaceous accreted terranes in Colombia, Ecuador, and Perú. Brown terranes are composed of continental crust and blue terranes are composed of oceanic crust (modified from: Aspden et al., 1987; Jaillard et al., 1990; Irving, 1975; Zamora and Litherland, 1993a; INGEMMET 1995 and 2000; Aleman and Ramos, 2000).

Within Ecuador the continental Chaucha terrane was accreted to the north of the Amotape-Tahuin terrane, and although the age of its accretion is not precisely known, it can be deduced that it was accreted before the Amotape-Tahuin accretion event (Zamora and Litherland, 1993a).

The Amaime terrane in Colombia is interpreted to be a piece of obducted oceanic crust accreted during the upper Middle Neocomian (Toussaint and Restrepo, 1982; Aspden et al., 1987). In Colombia, as in Ecuador and northwestern Perú, a westward shift of the volcanic arc occurred after accretion (Toussaint and Restrepo, 1982; Aspden et al., 1987; Mourier et al., 1988).

The accretion of the Amaime, Amotape-Tahuin, and Chaucha terranes ended the Vicusian orogeny (Toussaint and Restrepo, 1982; Aspden et al., 1987; Feininger, 1987; Mourier et al., 1988; Arculus et al., 1999). The volcanic arc shifted westward and the new subduction zone formed along the Cauca-Patia fault (Colombia), Pijili-Pallatanga fault (Ecuador), and an unidentified fault in Perú (Toussaint and Restrepo, 1982; Aspden et al., 1987; Feininger, 1987; Mourier et al., 1988; Zamora and Litherland, 1993a; Arculus et al., 1999; Fig. 2.4).

The difference in timing of the westward jump of the volcanic arc is important. The accretion of the Amaime and Chaucha terranes in Colombia and Ecuador may have occurred slightly earlier than the accretion of the Amotape-Tahuin terrane in Ecuador and Perú (Hall and Calle, 1982; Toussaint and Restrepo, 1982; Aspden et al., 1987; Jaillard et al., 1990; Zamora and Litherland, 1993a). This event stopped volcanic activity in central and northern Colombia, whereas in southernmost Colombia, Ecuador, and northernmost Perú, volcanic activity continued until the Amotape-Tahuin and Chaucha terranes were

accreted (Hall and Calle, 1982; Toussaint and Restrepo, 1982; Aspden et al., 1987; Jaillard et al., 1990; Zamora and Litherland, 1993a). This Late Jurassic to Early Cretaceous burst of volcanic activity deposited the Misahualli Unit of the Chapiza Fm. in Ecuador, and its equivalent in Perú, the Sarayaquillo Fm. (Tshopp, 1953; Kennerly, 1980; Baldock, 1982; Tafur, 1991; Zamora and Litherland, 1993a; Quispesivana, 1996; Palacios, 1997).

In Ecuador and Perú, a sedimentary hiatus occurred during the Neocomian (144-119 Ma) resulting in a major regional unconformity (Tshopp, 1953; Kennerly, 1980; Hall and Calle, 1982; Baldock, 1982; Tafur, 1991; Zamora and Litherland, 1993a; Quispesivana, 1996; Palacios, 1997). Subsidence throughout the region beginning in the Aptian saw the deposition first of sandstones derived from the east (named the Hollin Fm. in Ecuador, the Goyllarisquiza Group in Perú, and the Caballos Fm. in Colombia) and later, as transgression progressed, deposition of a thick shale sequence (the Napo Fm. in Ecuador and Chonta Fm in Perú; Macellari, 1988; Zamora and Litherland, 1993a).

Late Cretaceous-Paleocene

During the Late Cretaceous-Paleocene the Western Cordillera, Piñón, Piñón-Macuchi, and Talara Basin (?) terranes were accreted along the Cauca-Patia fault (Colombia), Pijili-Pallatanga fault (Ecuador), and an unidentified fault in Perú (Toussaint and Restrepo, 1982; Aspden et al., 1987; Feininger, 1987; Mourier et al., 1988; Arculus et al., 1999; Fig. 2.4). These terranes (with the exception of the Talara Basin) are dominated by basic igneous rocks that represent both oceanic crust and the remnants of an oceanic island-arc (Toussaint and Restrepo, 1982; Aspden et al., 1987; Feininger, 1987; Lebras et al., 1987; Mourier et al., 1988; Arculus et al., 1999). The Talara Basin is

composed of a thick sequence of Campanian to Late Eocene sedimentary rocks believed to have been derived from the west (Feininger, 1987). In Colombia, Ecuador, and northwestern Perú, the volcanic arc shifted westward at this time (Toussaint and Restrepo, 1982; Zamora and Litherland, 1993a).

3. Geologic history of the Cordillera del Condor

The Cordillera del Condor forms a chain of NNE trending mountains in the subandean border region of Ecuador and Perú (Fig. 1.1).

Precambrian

The basement rocks of the Cordillera del Condor consist of a thin belt of poorly exposed, north south striking, unnamed Precambrian (?) migmatitic gneiss in Ecuador (Zamora and Litherland, 1993a; Figs. 3.1 and 3.2). These metamorphic rocks are better exposed in northern and central Perú where they have been named the Complejo del Mara  n, or Mara  n complex, which consists of reddish gneissose phyllites and schists (Wilson et al., 1967; Quispesivana, 1996).

The metamorphic grade decreases in a southerly direction. Outcrops in the upper Mara  n River valley in north central Perú are comprised of green and gray micaceous chloritic schists, whereas further south in central Perú the complex grades into phyllites and gray slates (Wilson et al., 1967; Carlier et al., 1982; Quispesivana, 1996). Carlier et al. (1982) recognized two metamorphic and four deformation events in north central Perú that effected these Precambrian rocks. The age of the first metamorphic event is dated at 600 ± 50 Ma by the U-Pb method on metamorphic zircons (Dalmayrac, 1978, cited in Carlier et al., 1982). The age of the second event is poorly constrained but is believed to be pre-Carboniferous (Carlier et al., 1982). The first and second deformations coincide with their respective metamorphic events. The third and fourth deformations are presumed to be pre-Carboniferous but precise ages are not known (see the Pumbuiza Fm. below).

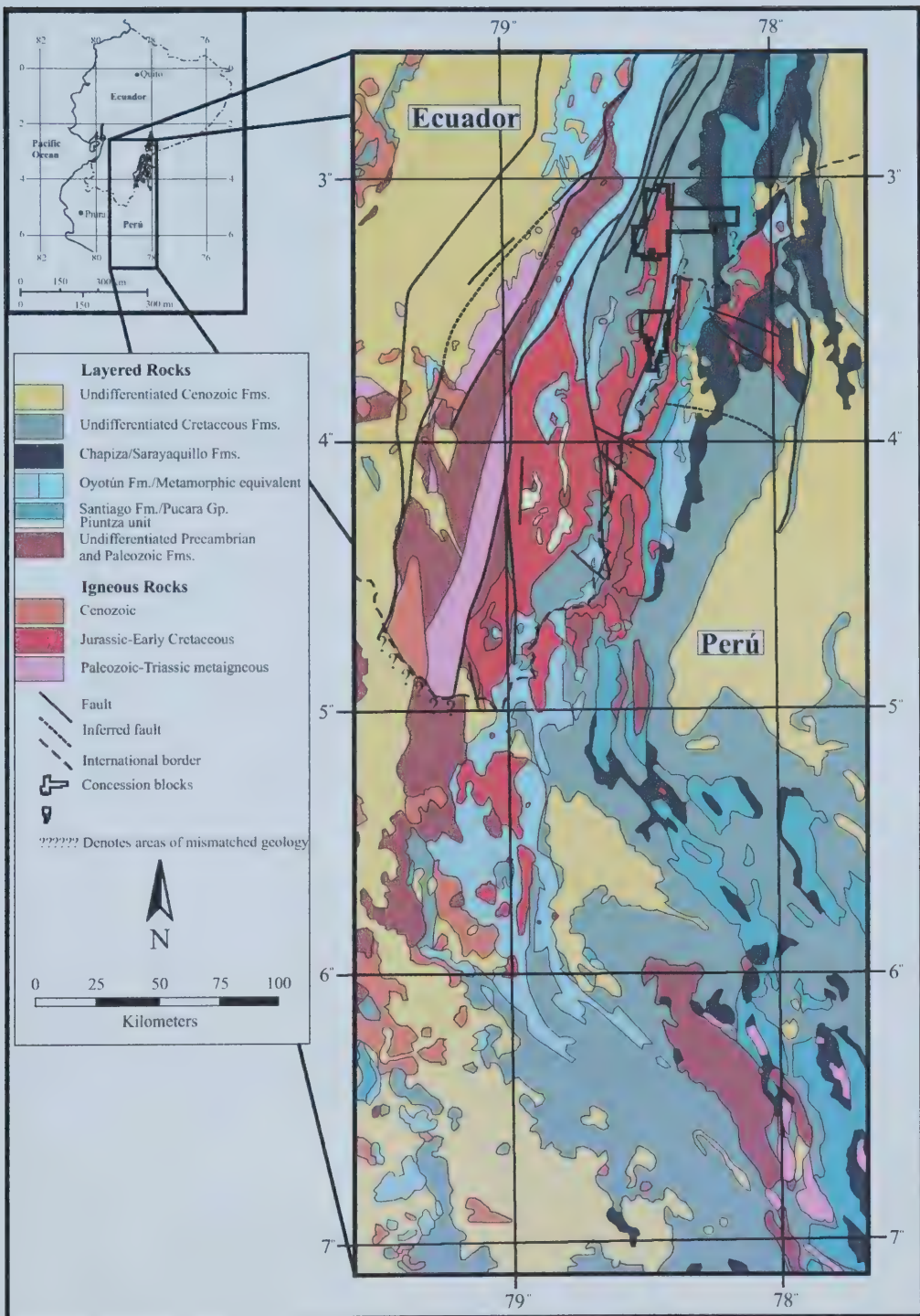


Fig. 3.1. Regional geologic interpretation map of the Cordillera del Condor region (adapted from Zamora and Litherland, 1993a; INGEMMET, 1995 and 2000).

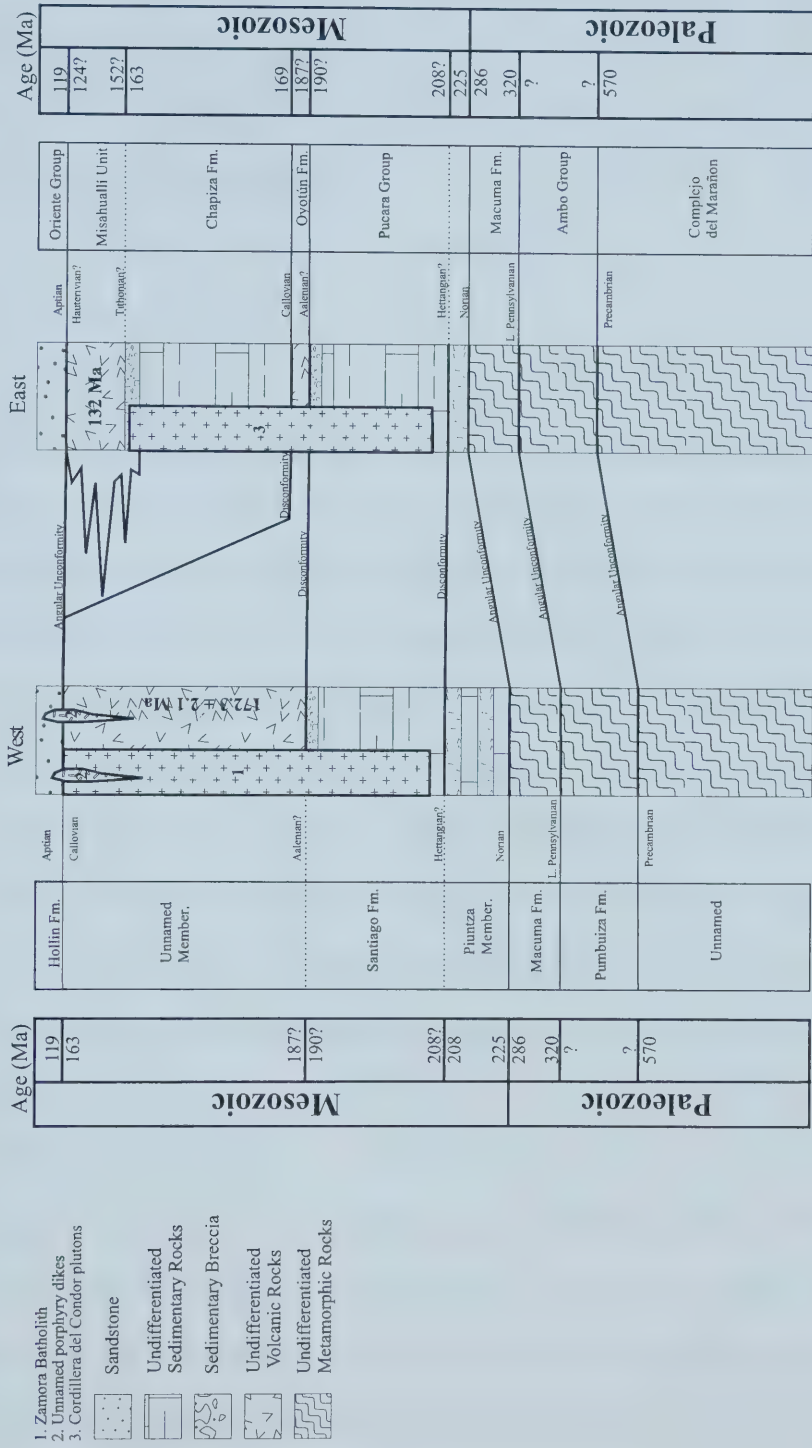


Fig. 3.2. Stratigraphic correlation chart between the Ecuadorian (western) and Peruvian (eastern) sides of the Cordillera del Condor (adapted from: Tschopp, 1953; Cobbing et al., 1981; Kennerly, 1980; Hall and Calle, 1982; Baldock, 1983; Wilson, 1984; Tafur, 1991; Zamora and Litherland, 1993a; Romeuf et al., 1995; Quispesivana, 1996; Palacios, 1997).

Small serpentized mafic to ultramafic bodies of Late Precambrian to Cambrian age intrude the Complejo del Mara  n (Wilson et al., 1967; Carlier et al., 1982), followed by a post-tectonic diorite-granodiorite-granite series which intruded prior to the Carboniferous (Carlier et al., 1982).

Paleozoic

Late Silurian-Mississippian

In Ecuador, the Precambrian rocks are unconformably overlain by the Pumbuiza Fm. (Fig. 3.2). This formation consists of a series of strongly deformed and faulted, dark gray to black, thin bedded slates, intercalated with quartzitic sandstones of unknown thickness (Tschopp, 1953; Baldock, 1982). The formation is only known to crop out in the area of Cerro Macuma, which is located within the Cordillera de Cutucu some 25km to the north-northeast of the Cordillera del Condor (Fig. 1.1; Tschopp, 1953; Zamora and Litherland, 1993a). Balkwill et al. (1995) noted that marine fossils from the formation provide ages ranging from the Late Silurian to Mississippian, and that the formation is overlain by the Early Pennsylvanian Macuma Fm., providing an upper age constraint.

The Ambo Group outcrops in north central Per   and appears to correlate with the Pumbuiza Fm. of Ecuador, although Quispesivana (1996) does not mention the Ambo Group as outcropping on the Peruvian side of the Cordillera del Condor. South of the Cordillera del Condor in the upper Mara  n River valley of Per  , Wilson et al. (1967) described the Ambo Group as a thick series of sandstones, shales, and conglomerates, which overlies the Complejo del Mara  n with an angular unconformity. Fossil evidence from strata at the base of the group gave a Devonian age, whereas the upper strata yielded a possible Mississippian age. Other authors have assigned a Mississippian age to the entire

Formation (e.g., Laubacher and Megard, 1985). Although the Ambo Group is time correlative with the Pumbuiza Fm. in Ecuador, it should be noted that the facies are different. The Ambo Group appears to be a proximal facies whereas the Pumbuiza Fm. has distal characteristics. This implies that the source region was located to the southeast in north-central Perú.

Pennsylvanian

The Late Carboniferous is represented in the Cordillera del Condor in Ecuador by sporadic outcrops of the Macuma Fm., which overlies the Pumbuiza Fm. in an apparent unconformity (Fig. 3.2; Tschopp, 1953; Zamora and Litherland, 1993a). The Macuma Fm. as shown on the geologic map of Ecuador (Zamora and Litherland, 1993a) consists of a shale and limestone sequence located at Cerro Macuma in the Cutucu Cordillera. Further south, in the Cordillera del Condor, the Macuma Fm. consists of metasedimentary rocks intercalated with metavolcanic rocks occurring as large roof pendants within the Jurassic Zamora Batholith (Zamora and Litherland, 1993a; McKelvey, 1991). No correlative lithologies have been identified from the Peruvian side of the Cordillera del Condor. Further westward in Ecuador, time equivalent metamorphic units crop out in the Cordillera Real and Amotape-Chaucha terrane (Zamora and Litherland, 1993a). Feininger (1987) states that the Macuma Fm. is Permian in age but does not corroborate this.

Permian-Mesozoic

Within the Cordillera del Condor three volcanic rock sequences have been observed (Figs. 3.1 and 3.2). The first sequence (the Piuntza Member of the Santiago Fm.) is dated as Late Triassic, and apparently corresponds to the terminal phase of the

Permo-Triassic rift (Zamora and Litherland, 1993a). It is possible that semi-continuous volcanism occurred throughout the time-span of the rift, and this first event may be related to the Late Permian volcanic strata in the Mitu Group from north-central Perú (Wilson et al., 1967). The second event (which has not been named) began during the Early Jurassic and continued until the upper Middle Jurassic, and provides the first evidence of subduction in the region during this period (Aspden et al., 1987; Litherland and Zamora, 1993a; Romeuf et al., 1995). The third event (the Misahualli Member of the Chapiza Fm.) preserves evidence of voluminous volcanic activity that commenced during the Late Jurassic and continued into the Early Cretaceous (Tschopp, 1953; Hall and Calle, 1982; Zamora and Litherland, 1993a).

The latter two volcanic sequences are related to the emplacement of a calc-alkaline batholith chain in the Eastern and Central Cordilleras of Colombia, the Cordillera Real of Ecuador, and the Western Cordillera of northern Perú (Tschopp, 1953; Toussaint and Restrepo, 1982; Aspden et al., 1987; Litherland and Zamora, 1993a; Quispesivana, 1996).

Permian-Early Triassic

Permian to Early Triassic lithologies are not known from the Cordillera del Condor (Tschopp, 1953; Zamora and Litherland, 1993a; Quispesivana, 1996).

Late Triassic-Early Jurassic

Late Triassic rocks outcrop on both sides of the Cordillera del Condor (Fig. 3.1). In Perú, the Pucara Group rests with an angular unconformity on the Precambrian Complejo del Mara  n (Tafur, 1991; Quispesivana, 1996). The Pucara Group consists of three formations: the Chambara, Aramachay, and the Condorsinga. Together, these

formations comprise an 800m thick package of dark colored limestones with thin intercalations of shales, rare tuffs, and calcareous breccias in the upper sections. Fossil dates from the middle strata range from the Early to Middle Liassic (Tafur, 1991; Quispesivana, 1996). The outcrop of the Pucara Group ranges from the latitude of Lima north to the south-central part of eastern Ecuador where it correlates with the Santiago Fm. (Tchopp, 1953; Cobbing et al., 1981; Tafur, 1991; Zamora and Litherland, 1993a).

Along the Ecuadorian side, the Piuntza Member of the Santiago Fm. crops out in association with the metamorphic rocks of the Carboniferous Macuma Fm. as large roof pendants in the Zamora Batholith (McKelvey, 1991; Zamora and Litherland, 1993a; Meinert, 1998). The Piuntza Unit is a 30 to 500m-thick subaqueous volcano-sedimentary sequence of andesitic to dacitic lavas, pyroclastic rocks, sandstone, shale, and limestone of Late Triassic age (Zamora and Litherland, 1993a; Meinert, 1998; Rodgers, 1998).

The Piuntza Member and the lower formation of the Pucara Group (Chambara Fm.) are time correlative. A westward facies change between the carbonate sequence of the lower Pucara Group and the volcano-sedimentary Piuntza Member was documented by Tafur (1991). He described the Carhuamayo Fm. (evidentially the Chambara Fm. of the Pucara Group), as an approximately 430m thick, shallow water limestone sequence containing minor intercalations of black shales and a tuff unit in the upper strata. The Carhuamayo/Chambara Fm. is overlain by Early Liassic limestones of the second Pucara Group formation, the Aramachay Fm., which places the minimum age of the Carhuamayo/Chambara Fm. below the base of the Sinemurian (Tafur, 1991).

Early-Late Middle Jurassic

The Pucara Group/Santiago Fm has been noted by a number of authors to change facies from dominantly sedimentary in the east to volcano-sedimentary in the west (Figs. 3.1 and 3.2; Tschopp, 1953; Cobbing et al., 1981; Kennerly, 1980; Baldock, 1982; Tafur, 1991; Palacios, 1997). The eastern facies of the Santiago Fm. was described by Tschopp (1953), and consists of thin bedded gray to black limestones alternating with sandstones, shales, minor tuffs, and tuffaceous shales. Tschopp (1953) observed a facies change westward from the Santiago River where the formation contained a greater volcanic component. Cobbing et al. (1981) observed the same facies change within the Pucara Group, where its western equivalent was denoted as the Zaña Group volcano-sedimentary sequence.

Within the Peruvian Cordillera del Condor, the Oyotún Fm. is described as gray to dark green aphanitic andesite lavas with rare dacite flows intercalated with volcanic breccias, tuffs, and sediments of an Late Liassic to late Middle Jurassic age (Quispesivana, 1996). The sequence rests disconformably on the upper strata of the Pucara Group (Quispesivana, 1996; Palacios, 1997). Palacios (1997) mentions that within the Western Cordillera of Perú the Oyotún Fm. rests on the Triassic carbonate La Leche Fm. and fossils from sediments within the Oyotún Fm. provide an age range from the Hettangian to Aalenian (205-180 Ma). Hyaloclastites, flow-dome complexes, and near shore sedimentary rocks described within the Oyotún Fm. indicate a shallow water subaqueous depositional environment (Quispesivana, 1996; Palacios, 1997).

In Ecuador, an equivalent volcanic sequence exists within the Santiago Fm., which was not given member status (Figs. 3.1 and 3.2; Tschopp, 1953; Baldock, 1982).

This sequence has been denoted as the Misahualli Member of the Santiago Fm. by Zamora and Litherland (1993a). Romeuf et al. (1995) observed two volcanic sequences in the field but did not clarify their relationship, choosing the lower sequence for their study, they denoted it as the Misahualli Fm., and obtained an Ar-Ar plateau age of 172 ± 2.1 Ma from an andesite.

The Misahualli Member as originally described by Tschopp (1953) is part of the Chapiza Fm., which consists of a red-bed sandstone and shale sequence with minor evaporites, capped by the volcano-sedimentary Misahualli Member which crops out mainly to the north of 3°S. Zamora and Litherland (1993a) combined the Chapiza Fm. into the Santiago Fm. This scheme is unworkable because it combines two separate tectonic regimes, a post rift sag basin (Santiago Fm./Pucara Gp.) and rocks related to subduction (Oyotún and Chapiza Fms.) which do not constitute a continuous geologic event (see below).

The Pucara Gp./Santiago Fm. was intruded by the Zamora Batholith and its Peruvian equivalent (the Cordillera del Condor plutonic body) beginning during the Early Jurassic (Figs. 3.1 and 3.2; Zamora and Litherland, 1993a; Quispesivana, 1996). The Zamora Batholith is a large igneous body that crops out over at least 1800km² in Ecuador (Hall and Calle, 1982). Granodiorite, granite, quartz monzonite, and syenite have been described from the Zamora Batholith (Kennerly, 1980; Baldock, 1982; Hall and Calle, 1982; Gemuts et al., 1992; McKelvey, 1991; McKelvey and Hammarstrom, 1991; Hammarstrom, 1993; Zamora and Litherland, 1993a). Two volcano-sedimentary sequences host the Zamora Batholith, the Late Triassic Piuntza Member and an unnamed Middle Jurassic Member of the Santiago Fm. (named the Oyotún Fm. in Perú; Tschopp,

1953; Kennerly, 1980; Baldock, 1982; McKelvey, 1991; Tafur, 1991; McKelvey and Hammarstrom, 1991; Hammarstrom, 1993; Gemuts et al., 1992; Zamora and Litherland, 1993a; Romeuf et al., 1995; Quispesivana, 1996).

Late Jurassic-Early Cretaceous

The upper Middle Jurassic was characterized by a period of uplift and regression along with a hiatus in volcanic activity (Tschopp, 1953; Cobbing et al., 1981; Kennerly, 1980; Baldock, 1982; Tafur, 1991). The Chapiza/Sarayaquillo Fms. began to be deposited as early as the Middle Jurassic but deposition continued into the Early (?) Cretaceous: they consist of a thick sequence of subaerial sandstones and shales containing minor evaporites in the lower section, capped by a variable sequence of volcanic rocks in Ecuador and Perú respectively. (Fig. 3.2; Tschopp, 1953; Cobbing et al., 1981; Kennerly, 1980; Baldock, 1982; Tafur, 1991). The lowest strata of the Chapiza/Sarayaquillo Fm. interfinger with the uppermost strata of the Santiago/Oyotún Fm. within Perú (Tafur, 1991). The source of the Chapiza Fm. was the volcanic arc, which shed sediments into the back-arc basin located to the east (Hall and Calle, 1982).

Intrusive activity apparently continued during this period with parts of the Zamora Batholith/Cordillera del Condor Plutonic Complex intruding the Chapiza/Sarayaquillo Fms. in Perú (Fig. 3.1; INGEMMET, 2000).

Late Early Cretaceous–Tertiary

A brief summary of the Late Early Cretaceous – Tertiary periods compiled from Tschopp (1953), Cobbing et al., (1981), Kennerly (1980), Baldock (1982), Zamora and Litherland (1993a) is given below.

During the Aptian-Albian the region rapidly subsided and a sandstone sourced from the east (the Hollin Fm. and Goyllarisquizga Gp.) was unconformably deposited over older rocks in Ecuador and Perú. During the Late Cretaceous the transgression continued depositing a marine shale unit over the sandstone. Uplift of the region commenced during the Paleocene and has continued throughout the Tertiary. Erosion related to this uplift has exposed Mesozoic and older rocks in the Cordillera del Condor and Cordillera de Cutucu in the last few million years.

4. Vicusian Metallogeny

Introduction

The Jurassic volcanic arc hosts the oldest well-documented, widespread, Mesozoic copper mineralization along the western margin of South America. The copper mineralization was emplaced during a period that spanned the upper Middle Jurassic-Early Cretaceous (Sillitoe et al., 1982, 1984; Espinoza et al., 1996; Marschik and Fontboté, 1996; Vila et al., 1996; Sillitoe, 1988; Gendall et al., 2000). During the Middle Cretaceous-Early Tertiary, localized overprinting of the copper belt by low sulfidation gold mineralization occurred within the southern part of the northern segment of the volcanic arc (e.g., the Nambija District; Gemuts et al., 1992; Zamora and Litherland, 1993b; Rodgers, 1998; Gendall et al., 2000).

The volcanic arc is divided into a northern and a southern segment. The northern segment spans Colombia, Ecuador, and northwestern Perú, and is cross-cut by the Andean cordillera, whereas the southern segment spans southern Perú-Chile and is located along the present day Pacific coast (Fig. 4.1).

Plate movements and migration of mineralization

The opening of the South Atlantic Ocean during the late Early Cretaceous caused a change in the motion of the South American plate. The plate began to pivot in a clockwise direction, causing the volcanic arc to migrate west in the northern segment and east in the southern segment (Fig. 4.1 Aspden et al., 1987; Sillitoe, 1988; Zamora and Litherland, 1993a; INGEMMET 1995, 2000). Geologic maps of Colombia, Ecuador, and Perú show a decreasing amount of westward displacement of the Cretaceous arc (from north to south) as the town of Huancabamba is approached, which appears to have acted



Fig. 4.1. Regional sketch map of northwest South America showing the location of the Jurassic, Cretaceous, Paleogene, and Neogene volcanic arcs in Colombia, Ecuador, Perú, and northern Chile. Note the Jurassic volcanic arc is cut by the younger volcanic arcs in the Huancabamba region and the order of the arcs is reversed, except for the Neogene arc (modified from: Aspdén et al., 1987; Jaillard et al., 1990; Irving, 1975; Zamora and Litherland, 1993a; INGEMMET, 1995 and 2000; Noble and McKee, 1999; Aleman and Ramos, 2000).

as the pivot point (Irving, 1975; Mourier et al, 1988; Zamora and Litherland, 1993a; INGEMMET 1995, 2000). A gap in the Cretaceous volcanic arc appears to have existed in southern Colombia and throughout most of Ecuador (Aspden et al., 1987; Zamora and Litherland, 1993a). Whether this apparent arc gap is due to lack of preservation or non-deposition it is not known. The geology of the region of Huancabamba shows that the two Mesozoic volcanic arcs and their associated mineralization cross in this region (Mourier et al, 1988; Zamora and Litherland, 1993a; INGEMMET 1995, 2000).

The westward shift in intrusive activity and copper mineralization continued into the Paleogene (Fig. 4.1; Sillitoe, 1988; Zamora and Litherland, 1993a, 1993b; Gendall et al., 2000; INGEMMET 1995, 2000). The volcanic arc and associated mineralization then migrated east during the Neogene in Colombia (Aspden et al., 1987; Sillitoe, 1988). However, in Ecuador, the arc sporadically outcrops along the western edge of the Piñon-Macuchi terrane and only as Huancabamba is approached does the arc turn toward the southeast, crossing the older arcs (Fig. 4.1; Zamora and Litherland, 1993a; Gendall et al., 2000; Noble and McKee, 1999; INGEMMET 1995, 2000). The Yanacocha high sulfidation district is of Miocene age and is located within the Miocene volcanic arc to the southeast of the Nambija district (Fig. 4.2; Noble and McKee, 1999). The separation between the Pangui and Yanacocha districts is approximately 400km and each younger district appears to have been emplaced to the south of the older district (Fig. 4.2; Gendall et al., 2000).

Jurassic - Cretaceous mineralization

The porphyry copper deposits from the northern segment of the volcanic arc are the primary focus of this study. The southern segment of the volcanic arc hosts vein,

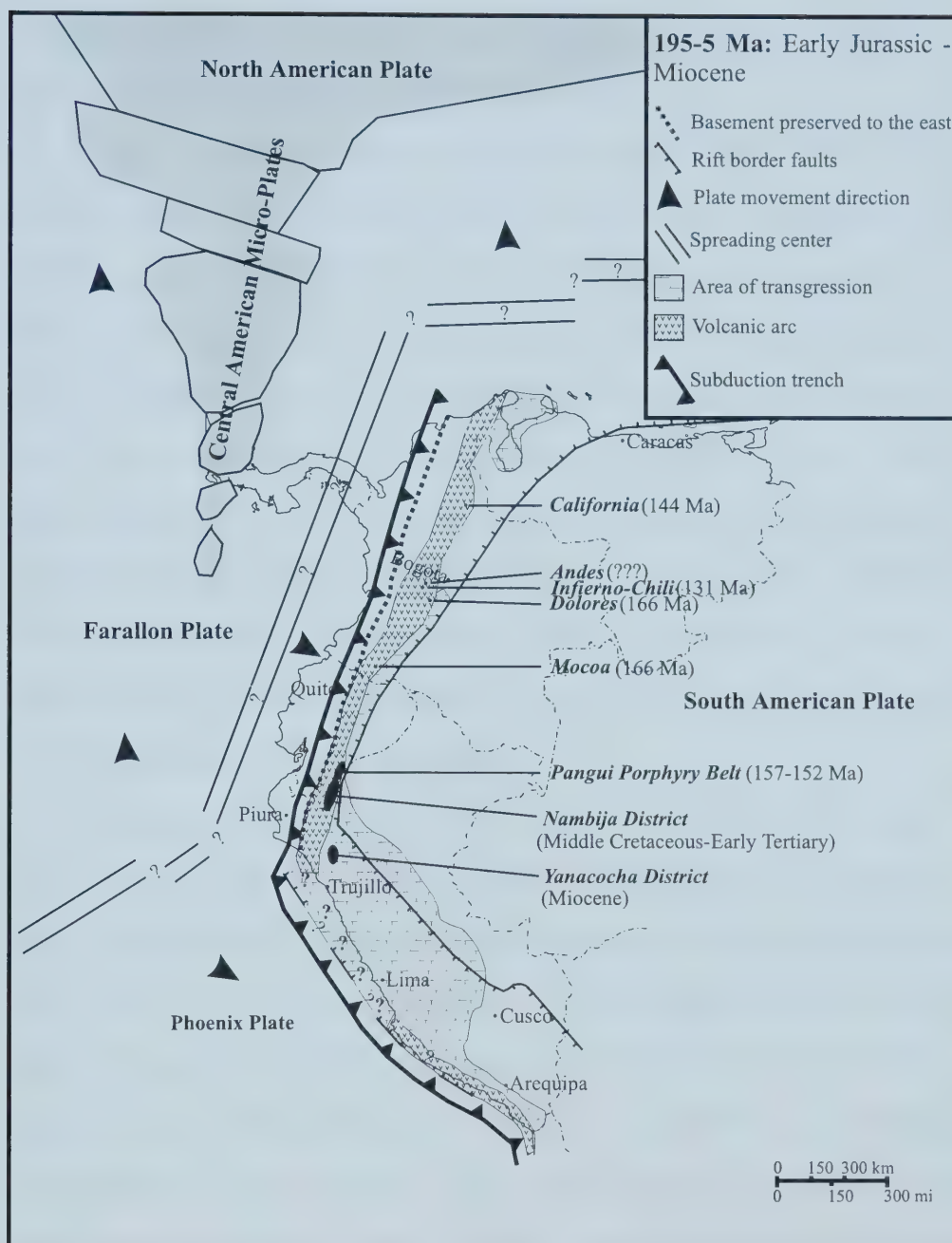


Fig. 4.2. Regional tectonic sketch map of northwest South America showing the location and age (in brackets) of porphyry prospects and the Pangui and Nambija Districts (Ecuador) and the Yanacocha District (Perú) (modified from: Aspdén et al., 1987; Jaillard et al., 1990; Irving, 1975; Sillitoe et al., 1982; Sillitoe, 1988; Zamora and Litherland, 1993a; INGEMMET, 1995 and 2000; Aleman and Ramos, 2000).

manto, breccia-pipe, and minor porphyry deposits and will only be discussed briefly (Vila et al., 1996; Marschik and Fontboté, 1996; Sillitoe, 1988).

Copper mineralization within the northern and southern segments of the volcanic arc spanned a period of approximately 174-128 Ma (Sillitoe et al., 1982; Espinoza et al., 1996; Vila et al., 1996; Marschik and Fontboté, 1996; Sillitoe, 1988; Gendall et al., 2000). Porphyry copper deposits of this age are known from Colombia (e.g., Mocoa, Dolores, California, Infierno-Chili, and Andes; Fig. 4.2) and from Ecuador, where the Pangui district contains 11 known prospects (Sillitoe et al., 1982; Gemuts et al., 1992; Zamora and Litherland, 1993b; Gendall et al., 2000; Fig. 4.3; Table 4.1). Three periods of porphyry copper mineralization corresponding to 166 ± 4 , 144 ± 3 , and 131 ± 2 Ma were observed by Sillitoe et al. (1982) in Colombia. Within the Pangui District of Ecuador, radiometric age-dating shows that two periods of mineralization occurred, which spanned 157-152 Ma (Gendall et al., 2000; this study).

The age and composition of the host rocks for the deposits vary from Late Triassic to Early Cretaceous sedimentary, volcanic, and intrusive rocks (Sillitoe et al., 1982, 1984; Gendall et al., 2000). All of the deposits appear to be related to the late stages of large batholiths and also appear to have been emplaced semi-continuously during the upper Middle Jurassic-Lower Cretaceous (Sillitoe et al., 1982, 1984; Sillitoe, 1988; Gendall et al., 2000). During the emplacement of the batholiths, minor base metal skarn formation occurred (Sillitoe et al., 1982, 1984; McKelvey, 1991; McKelvey and Hammarstrom, 1991; Hammarstrom, 1993; Gemuts et al., 1992; Litherland et al., 1992; Gendall et al., 2000). There is some confusion in the literature about the deposit types within the Nambija gold district (e.g., gold skarn vs. epithermal; Meinert, 1998).



Fig. 4.3. Sketch map showing the concessions and location of the individual prospects.

Table 4.1 Deposits located within the Jurassic-Early Cretaceous volcanic arc

Deposit name	Age	Metals	Host Fm.	Classification	Reference(s)
Colombia					
California	J	Cu-Mo-Au?		Porphyry	
Dolores	J	Cu-Mo		Porphyry	
Inferno-Chili	J/K	Cu-Mo		Porphyry	Sillitoe et al., 1982, 1984; Sillitoe, 1988
Mocoa	J	Cu-Mo		Porphyry	
Ecuador, Pangui District					
Chancho	J	Cu-Mo-Au	Zamora Batholith	Porphyry	
Kunucus	J	Cu	Santiago/Oyotún Fm.	Skarn	
La Florida	J	Cu-Mo	Zamora Batholith	Porphyry	
Panantz	J	Cu-Mo-Au	Zamora Batholith	Porphyry	
San Carlos	J	Cu-Mo	Zamora Batholith	Porphyry	
San Luis	J	Cu-Mo	Zamora Batholith	Porphyry	Gendall et al., 2000
San Marcos	J	Cu-Mo	Zamora Batholith	Porphyry	
San Miguel	J	Cu-Mo	Zamora Batholith	Porphyry	
Sutsu	J	Cu-Mo	Zamora Batholith	Porphyry	
Warintza	J	Cu-Mo	Santiago/Oyotún Fm.	Porphyry	
Wawarne/Mirador	J	Cu-Mo-Au	Zamora Batholith	Porphyry	
Ecuador, Nambija District					
Augusta	T	Au-poly	Napo Fm.	Epithermal	
Cambana-Campanilla	J?	Au	Santiago Fm. (Puntza unit)	Skarn	
Cerro Quemado	T	Au-poly	Napo Fm.	Epithermal	
Chinapintza	T	Au-poly	Eocene porphyry	Epithermal	
Conguine	J	Cu-Mo	PC Migmatitic Gneiss	Skarn	
Cunay-Fortuna	J?	Au	Santiago Fm. (Puntza unit)	Skarn	
El Hito	J?	Au	Macuma Fm.	Epithermal	Zamora and Litherland, 1993b
Guaysimi	J?	Au	Santiago Fm. (Puntza unit)	Skarn	
Jardín del Condor	J?	Au-poly	Misahualli Fm.	Epithermal	
La Esperanza	J	Cu-Mo	Santiago Fm. (Puntza unit)	Epithermal?	
Los Planes	J?	Au	Misahualli Fm.	Epithermal	
Maria Elena	J	Fe-Cu-Au	Santiago Fm. (Puntza unit)	Skarn	
Nambija	K	Au	Santiago Fm. (Puntza unit)	Skarn	Zamora and Litherland, 1993b; Gendall et al., 2000
Pangui / Biche	T	Au-poly	Macuma Fm.	Epithermal	
Puntza	J?	Au-poly	Santiago Fm. (Puntza unit)	Epithermal	
Sangola	T	Au-Cu	Misahualli Fm.	Epithermal	
Shamatata	J?	Au-poly	Napo Fm?	Epithermal	
Sultana del Condor	J?	Au	Santiago Fm. (Puntza unit)	Skarn	Zamora and Litherland, 1993b
Tres Cerrios	T	Au	Zamora Batholith	Epithermal	
Yantzaza	J?	Au-poly	Misahualli Fm.	Epithermal	
Zumbi	J	Fe-Cu-Au	Santiago Fm. (Puntza unit)	Skarn	
Northern Perú					
Cañarico	T	Cu-Mo-Pb-Zn	Calipuy Gp.	Porphyry/Vein	Chacón et al, 1997; Cobbing et al, 1981
Cerro Las Palomitas	J?	?	Oyotún Fm.	?	
Cerros Alumbrai	J?	?	Oyotún Fm.	?	Chacón et al, 1997
Coyota	J?	Au, Cu, Pb	Oyotún Fm.	Porphyry?	
Culebras	J?	Pb-Zn?	Oyotún Fm.	?	Chacón et al, 1997; Atsushi et al, 1998
Ehuamarca	T	Cu-Mo?	Calipuy Gp.	Porphyry?	Chacón et al, 1997
Hualtan	J?	Au	Oyotún Fm.	Porphyry?	Chacón et al, 1997; Steinmüller and Chacón, 1996
Ia Granja	T	Cu-Mo-Pb-Zn	Calipuy Gp.	Porphyry/Vein	Cobbing et al, 1981; Chacón et al, 1997; Gendall et al., 2000
La Huaca / La Vega	T	Cu-Mo-Pb-Zn-Ag	Calipuy Gp.?	Porphyry/Vein? Skarn?	Cobbing et al, 1981; Petersen and Vidal, 1983
Las Huacillas	J?	Au	Oyotún Fm.	Porphyry?	Chacón et al, 1997; Steinmüller and Chacón, 1996
Limón	J?	Au	Oyotún Fm.	?	Chacón et al, 1997
Linderos	J?	Cu-Ag-Pb-Zn-Ar-Au	Oyotún Fm.	Breccia Pipe?	Palacios, 1997
Los Blancos	J?	?	Oyotún Fm.	Porphyry?	
Mercedes	J?	?	Oyotún Fm.	?	
Nuevo Porvenir	?	?	?	?	Chacón et al, 1997
Nuevo Trujillo	J?	Au	Oyotún fm.	Porphyry?	
Parano	T?	Cu-Mo	Calipuy Gp.?	Porphyry	Cobbing et al, 1981
San Ignacio III	C?	Au	Oyotún fm. and Goyllarisquiza Fm.	Paleoplacer-manto?	
Sta. Martha	J?	?	Oyotún Fm.	?	Chacón et al, 1997
Turnalina	T?	Cu-Mo-W?	Tertiary intrusive?	Breccia Pipe	Chacón et al, 1997; Cobbing et al, 1981
Unión	J?	?	Oyotún Fm.	?	Chacón et al, 1997
Coast of Chile					
Gatico	K	Cu		Porphyry	
Puntillas	J	Cu		Porphyry	Sillitoe, 1988

Notes to Table 4.1

T: Tertiary

C: Cretaceous

J: Jurassic

Radiometric ages cited by Gendall et al. (2000) show that the Nambija district is Middle Cretaceous and younger in age. Field relationships consistently describe the gold mineralization as cross-cutting subeconomic base metal skarns related to the Middle Jurassic-Early Cretaceous intrusion of batholiths (McKelvey, 1991; Gemuts et al., 1992; Litherland et al., 1992; Rodgers, 1998; Gendall et al., 2000)

The Middle Jurassic-Early Cretaceous porphyry copper deposits vary from Cu-Mo to Cu-Mo-Au. Two deposits in Colombia contain anomalous amounts of molybdenite and gold; Mocoa and California respectively (Sillitoe et al., 1982, 1984). Multiple mineralization events are observed for the deposits both within the Pangui district and in Colombia (Sillitoe et al., 1982, 1984; Blair, 2000; Gendall et al., 2000). The best-developed porphyry deposits generally contain high hypogene copper grades ($>0.5\%$ Cu;) and a number of the deposits host mineralized breccias (e.g., Mocoa, San Carlos, and Mirador) which locally constitute ore (Sillitoe et al., 1982, 1984; Blair, 2000; Gendall et al., 2000).

The deposits, particularly within the Pangui District, are characterized by disseminated copper mineralization: stockwork mineralization is not well developed. The lack of preservation of advanced argillic alteration at the prospects may indicate that erosion of the deposits has been moderate to deep. Within the Pangui district, the porphyry prospects are invariably located in close proximity to the contact of the Zamora Batholith and the enclosing host rocks (usually the volcanic facies of the Santiago Fm.).

Alteration

Alteration developed within the porphyry deposits consists of potassic, propylitic, phyllic, and intermediate argillic, with a single example of advanced argillic alteration at

the California deposit in Colombia (Sillitoe et al., 1982). The least developed porphyries (e.g., Infierno-Chili and Dolores) show weak potassic alteration with a broad propylitic halo and local overprinting by phyllic alteration (Sillitoe et al., 1982). Within the Pangui District, the alteration pattern shows early potassic and propylitic development, followed by a locally pervasive phyllic overprint when the porphyry deposits are exposed at high levels (Blair, 2000; Gendall et al., 2000). The same pattern was observed by Sillitoe et al. (1982, 1984) at the Mocoa prospect in Colombia. In many of the deposits (e.g., San Carlos, Panantza, Chanco, and Wawame/Mirador) a structurally controlled intermediate argillic overprint exists (Blair, 2000; Gendall et al., 2000).

Cretaceous mineralization

Terrane accretion in Colombia during the Early Cretaceous caused a westward shift in the volcanic arc and mineralization (e.g., the Infierno-Chili deposit in Colombia, dated at 131 Ma; Sillitoe et al., 1982; Toussaint and Restrepo, 1982; Arculus et al., 1999). In Ecuador, terrane accretion also occurred during the Early Cretaceous, although it appears to have occurred slightly later than the accretion in Colombia, which is consistent with the southward younging trend in the porphyry copper deposits (Aspden et al., 1987; Feininger, 1987; Mourier et al., 1988; Sillitoe, 1988; Zamora and Litherland, 1993a, 1993b; Arculus et al., 1999; Gendall et al., 2000). The shift in the volcanic arc to the west in Colombia was pronounced, whereas in Ecuador, the westward shift was not as great, although sparse outcrop may indicate a gap in the volcanic arc in Ecuador (Zamora and Litherland, 1993a). Because of this, the Cretaceous volcanic arc in Ecuador and north-westernmost Perú partially overlaps the Jurassic volcanic arc (Figs. 4.1 and 4.2). This

overlap has created a local superposition of two mineral districts (Pangui copper porphyry and Nambija low sulfidation gold) in the area of the Zamora Batholith.

The fact that the Nambija gold district is younger than and overprints the Pangui copper district has now been confirmed (Gendall et al., 2000). This Middle Cretaceous-Early Tertiary gold mineralization overprints subeconomic Jurassic base metal skarn mineralization (Gemuts et al., 1992; Zamora and Litherland, 1993b; Rodgers, 1998).

The Nambija District

During the period 1549 to 1599, a large portion of the Spanish colonial gold came from Ecuador. The locations of the mining districts of Zamora, Zaruma, Sevilla de Oro, and Logroña del Oro were documented in colonial records (Miller and Singewald, 1919). The Zamora (Nambija) gold district is hosted within the Zamora Batholith.

In 1549 the mining towns of Zamora, Zaruma (Portovelo), Sevilla de Oro, and Logroña del Oro (Cañar) were founded, but by 1599 a native uprising caused the abandonment of most of the towns. Although the town of Zamora was later resettled, the mines were forgotten (Miller and Singewald, 1919). Portovelo, near the town of Zaruma, apparently was the only mine to survive the uprising. Miller and Singewald (1919) described gold as being mined from the Nambija district, but were unable to classify the deposit type. Since the rediscovery of the mines in 1980, a few descriptions of the district have been published in the literature (e.g., McKelvey, 1991; McKelvey and Hammarstrom, 1991; Hammarstrom, 1993; Gemuts et al., 1992; Litherland et al., 1992; Meinert, 1998).

There are at least 15 prospects in the vicinity of the town of Nambija (Table. 4.1). The deposits were initially classified as skarn deposits. Some of the prospects appear to

be hosted by alkalic phases of the Zamora Batholith in which a sequence of skarn and hornfels altered sedimentary and volcanic rocks (the Piuntza Member of the Santiago Fm.) forms a roof pendant within the Zamora Batholith and are spatially related to an altered syenite (McKelvey, 1991; Zamora and Litherland, 1993a, 1993b). McKelvey (1991) states that the syenite (which has been altered to quartz, sericite, clays, and pyrite) is associated with granodiorite and tonalite. The mineralization is stated to be located in a structural breccia zone that crosscuts a complex calcic skarn assemblage developed within the xenolith (McKelvey, 1991; McKelvey and Hammarstrom, 1991).

Aspden and Litherland (1992) described the Zamora Batholith as hosting: “Feldspar microporphyritic andesites, hornblende andesites, and dacites ... a series of small high level, subvolcanic intrusions. Some of the latter intrusions are associated with polymetallic Au mineralization and it is probable that they relate to a younger (post batholith) phase of activity (Aspden and Litherland, 1992, p. 198)”. Gemuts et al. (1992) also noted that at Nambija, “the source of the gold may be younger, perhaps Tertiary intrusives or the district may be related to a series of diatremes (Gemuts et al., 1992, p. 13)”. They also note that other porphyry-gold-type systems occur at Cutuntza, El Hito, and Augusta.

5. Description of the San Carlos and Panantza prospects

Introduction

Soil geochemistry, mapping, and drilling have delineated the dimensions of the two porphyry copper prospects at San Carlos and Panantza as approximately 2 x 2km and 1 x 2km respectively (see map insert).

The porphyry prospects are hosted by the Zamora Batholith, which in the prospect area consists of granite and granodiorite intruded into a Middle Jurassic volcano-sedimentary sequence (Gendall et al., 2000). The volcano-sedimentary sequence has not been named in Ecuador, but it is named the Oyotún Fm. in Perú (Tschopp, 1953; Tafur, 1991; Quispesivana, 1996). Radiometric age dating has identified two pulses of porphyry mineralization, the first occurred around 157 Ma and the second around 153 Ma (this study). The Zamora Batholith has been dated at approximately 160 Ma (this study).

The structural history of the Pangui region can be broken down into two main structural regimes: a Permian-Triassic rift, related to the breakup of Pangea into Gondwana and Laurasia, and early Jurassic to Holocene subduction (Irving, 1975; Aspden et al., 1987; Jaillard et al., 1990; Zamora and Litherland, 1993; Vila et al., 1996; Marschik and Fontbote, 1996; Benavides-Cáceres, 1999). Major rift related structures strike NNE-SSW and dextral movement along these major structures during the Jurassic developed secondary faults that strike NW-SE (Zamora and Litherland, 1993a, 1993b; Gendall et al., 2000).

The emplacement of the Zamora Batholith during the Middle (?) to Late Jurassic period was localized by the major rift related structures (Hall and Calle, 1982; Jaillard et al., 1990; Zamora and Litherland, 1993a; Balkwill et al., 1995). During or immediately

following the porphyry mineralization the region was uplifted and eroded (Tschopp, 1953; Hall and Calle, 1982; Zamora and Litherland, 1993a, 1993b). The sequence of intrusive and volcanic rocks is locally capped by dark green shale interpreted to be part of the lower Chapiza Fm. The entire region is unconformably overlain by the Hollín Fm., a cross-bedded sandstone of Early Cretaceous age (Tschopp, 1953; Zamora and Litherland, 1993a).

Prospect Geology

Oyotún Fm.-Chapiza Fm.

The Oyotún Fm. crops out on the periphery of both prospects and is composed of a dark gray fine-grained micro-porphyritic andesite, porphyritic dacite to rhyolite, and subordinate tuffs and flows (the tuffs and flows crop out between the San Carlos and Kutucus prospects). The dominant rock type in this formation is the dark colored micro-porphyritic andesite, which crops out as a very thick (up to 600m) mass that appears to form a dome. The rock does not contain flow structures and breaks into small rectangular blocks due to multiple joint sets. In thin section this rock consists of thin euhedral plagioclase laths in an aphanitic groundmass (devitrified glass?) with accessory magnetite (up to 5%) and micro-phenocrysts of what appear to have been pyroxene (Fig. 5.1). In the vicinity of the prospects the groundmass of this rock is moderately altered to carbonate $\pm$ sericite and it maintains its former textures. Trace amounts of pyrite were observed but no other mineralization.

The largest dacite and rhyolite porphyry flows crop out along the western edge of the Panantza prospect. The flows rest upon the micro-porphyritic andesite and although poorly exposed, they appear nearly flat lying and rather thin (up to 20m apparent

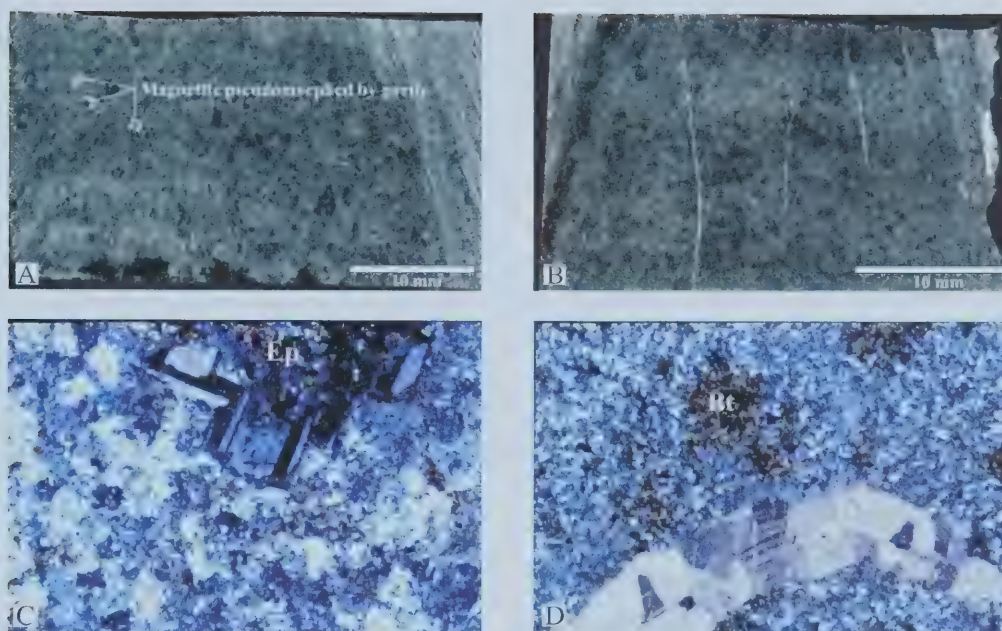


Fig. 5.1. Hand samples and thin sections of the Oyotún Fm. A. Hand sample showing preserved texture. The magnetite has been pseudomorphed by pyrite. B. Hand sample showing moderately pervasive secondary biotite and carbonate veinlets. C. Thin section showing alteration of groundmass to sericite. Observe the epidote (Ep) in the top center overprinting the relict feldspar (XPL, FOV, length of photo = 3.35mm). D. Thin section of B showing detail of the secondary biotite (Bt), pervasive sericite, and carbonate veinlets (XPL, FOV, length of photo = 3.35mm).

Thickness). In hand sample these rocks contain phenocrysts of feldspar and quartz in an aphanitic groundmass. These units do not appear to be altered or mineralized. The tuffs, flows, and shale (lower Chapiza Fm.) are exposed between the San Carlos and Kutucus prospects overlying the micro-porphyritic andesite with the shale cropping out in the upper part of the section. Dominantly white and buff colored tuffs form thin (to 3m) units with rare calcareous breccias separating the units. Rare, thin porphyritic (to 3m) dacite and rhyolite flows are interbedded with the tuffs and in appearance are very similar to the units previously described. The shale consists of an unfossiliferous orange-brown weathering dark green-black rock that forms a number of ridges. Around the San Carlos prospect area the tuff and flow unit is variably altered and mineralized. Skarn alteration containing rare garnets, pyrite, and pyrrhotite were observed but no other mineralization.

The lower contact with the Early Jurassic Santiago Fm. (a carbonate sequence) was not observed in the field. The Oyotún-Chapiza Fms. are unconformably overlain by the Hollin Fm. an Aptian-Albian sandstone (Tschopp, 1953). This major regional unconformity has also been observed in Perú (Tafur, 1991; Quispesivana, 1996).

Zamora Batholith

The Zamora Batholith covers an area of at least 2500 km² and is composed of a number of phases that include granite, granodiorite, syenite, and monzonite (Hall and Calle, 1982; McKelvey, 1991; McKelvey and Hammarstrom, 1991, 1993; Gemuts et al., 1992; Litherland et al., 1992; Mienert, 1998). Within the study area, the largest intrusive body is a rock that varies between granite and granodiorite. Field mapping has defined two phases of this body. The first and most widespread phase consists of a pink colored, equigranular, small-to-medium-grained, locally slightly porphyritic, hornblende-biotite-plagioclase granodiorite, containing interstitial microcline and quartz (Fig. 5.2). This phase outcrops on the periphery of the San Carlos prospect and is the dominant host rock of the Panantza prospect (see map insert). It covers an area of at least 5 x 10km. A second phase comprised of a white colored, porphyritic, inequigranular, medium-to-coarse-grained, hornblende-biotite granodiorite forms the core of the San Carlos prospect, along with a minor satellite body cropping out to the SE of the Panantza prospect. The larger body forms a rhombohedron of about 1.25 x 1.25km. Both granodiorite phases locally grade into granite.

The contact between the pink granodiorite and the Oyotún Fm. shows that the Zamora Batholith stopped into the Oyotún Fm. and the current level of erosion in the area has only just exposed the top of the Zamora Batholith (Fig.5.3).

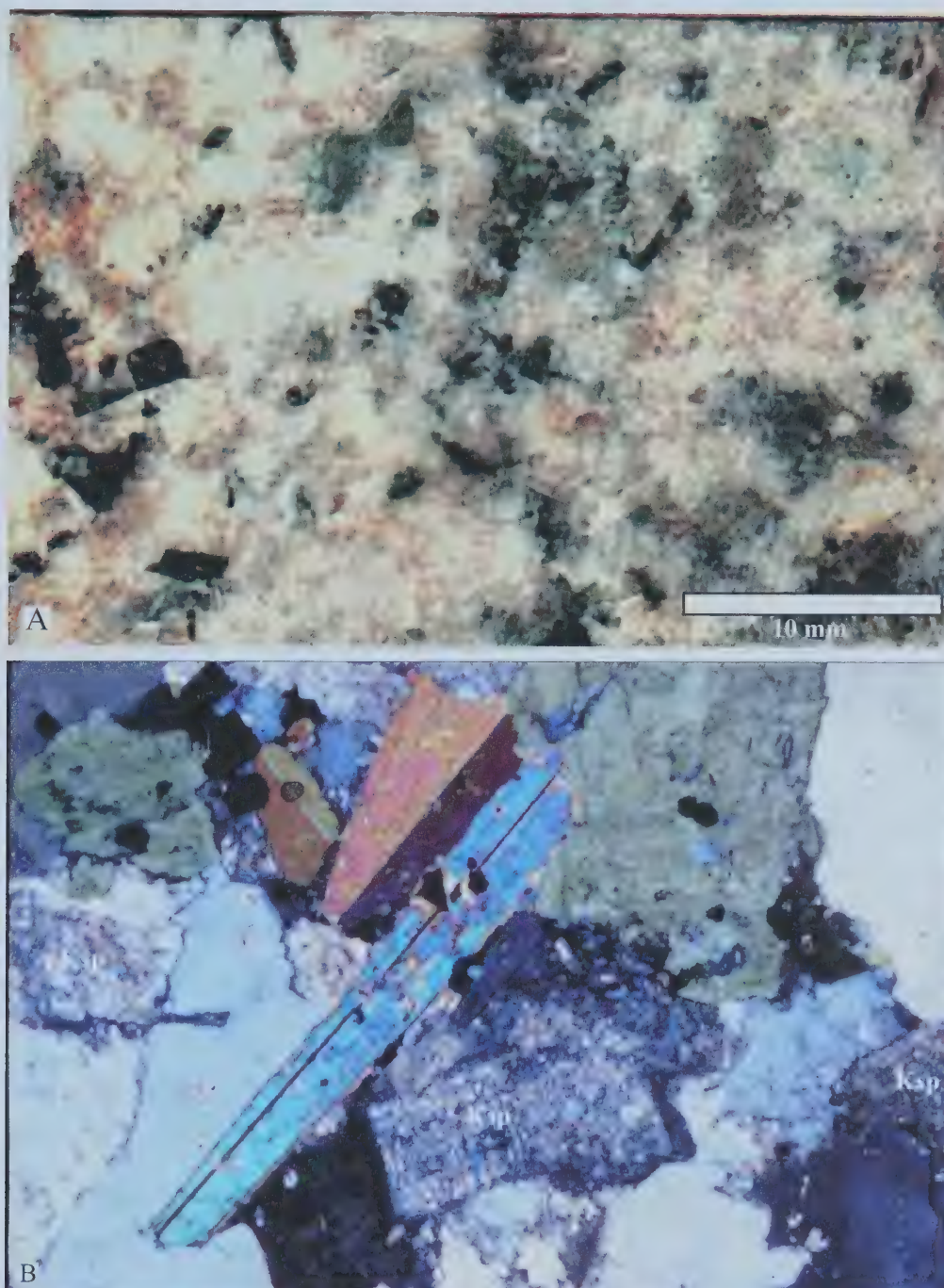


Fig. 5.2. Pink Zamora Batholith. A. Hand sample showing equigranular texture, abundant hornblende phenocrysts (black), and faint pink feldspars. B. Thin section showing essentially unaltered hornblende crystals and minor chlorite alteration of some mafic phases. The K-feldspar crystals (Ksp) have been selectively altered to sericite. FOV, length of photo $\approx 3.35\text{mm}$, XPL.



Fig. 5.3. Photograph taken in the Rio Akerones at the San Carlos prospect showing the stoped contact between the pink colored granodiorite of the Zamora Batholith and the dark green-gray andesite of the Oyotún Fm.

At the San Carlos prospect, a granite intrusion crops out to the east of the main center of mineralization. This granite is a light tan to flesh colored, aphanitic to pegmatitic rock containing perthitic feldspars and quartz, with a locally well developed graphic texture. This unit appears to be an irregularly shaped body about 500m in diameter (Fig. 5.4).

Porphyry dike relationships

Both prospects host two generations of similar composition mineralized porphyry dikes which are cross-cut by a later generation of small unmineralized, intermediate to felsic composition porphyry dikes.

Early porphyry dikes

At both prospects the early porphyry dikes consists of medium-grained, intermediate composition, hornblende-plagioclase (quartz monzodiorite to granite) porphyry (Fig. 5.5).

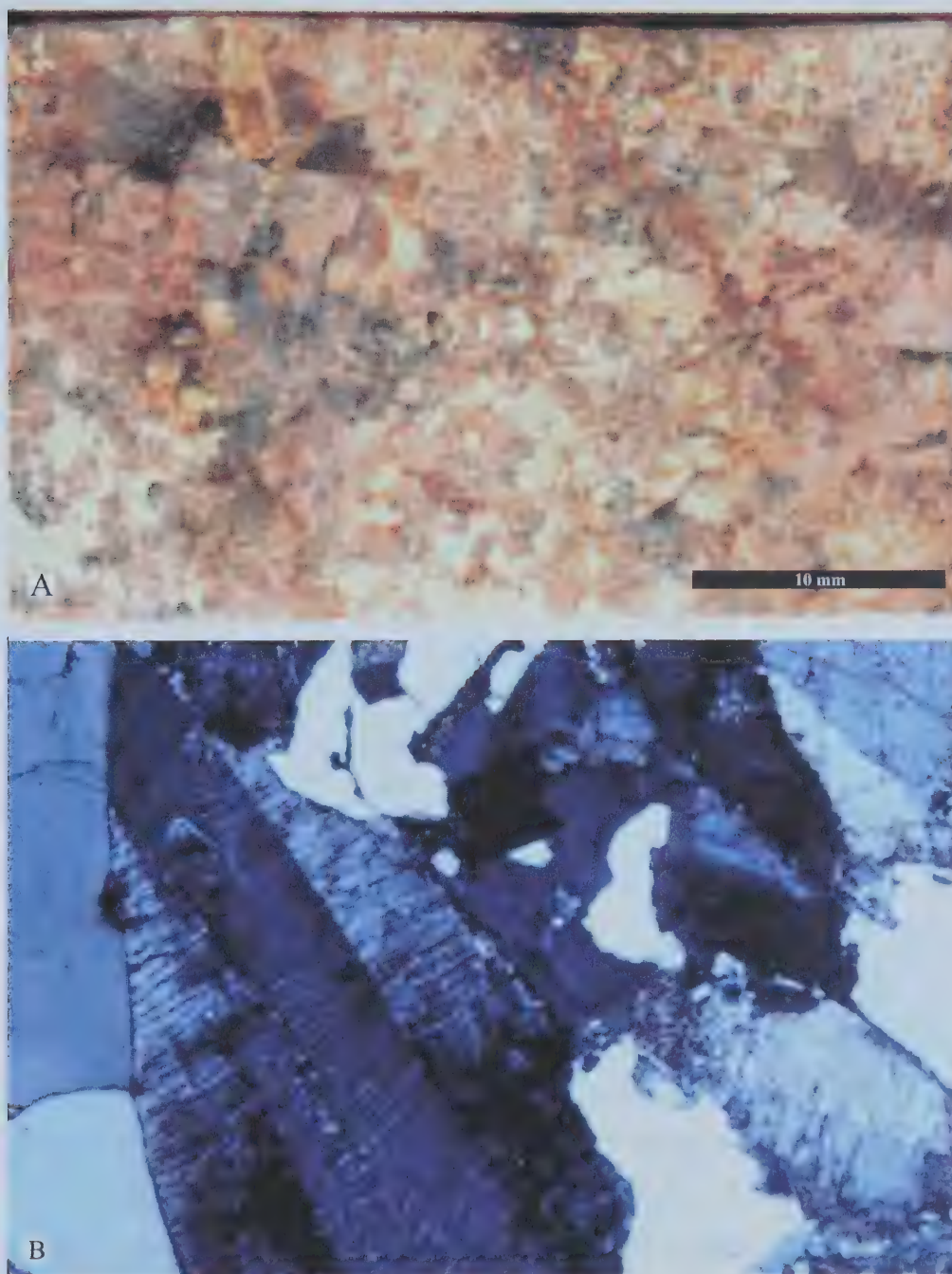


Fig. 5.4. Leucocratic granite of the Zamora Batholith from the San Carlos prospect. A. Hand sample showing the textural variations commonly observed. B. Thin section showing perthitic microcline and quartz (FOV, length of photo = 3.35mm, XPL).

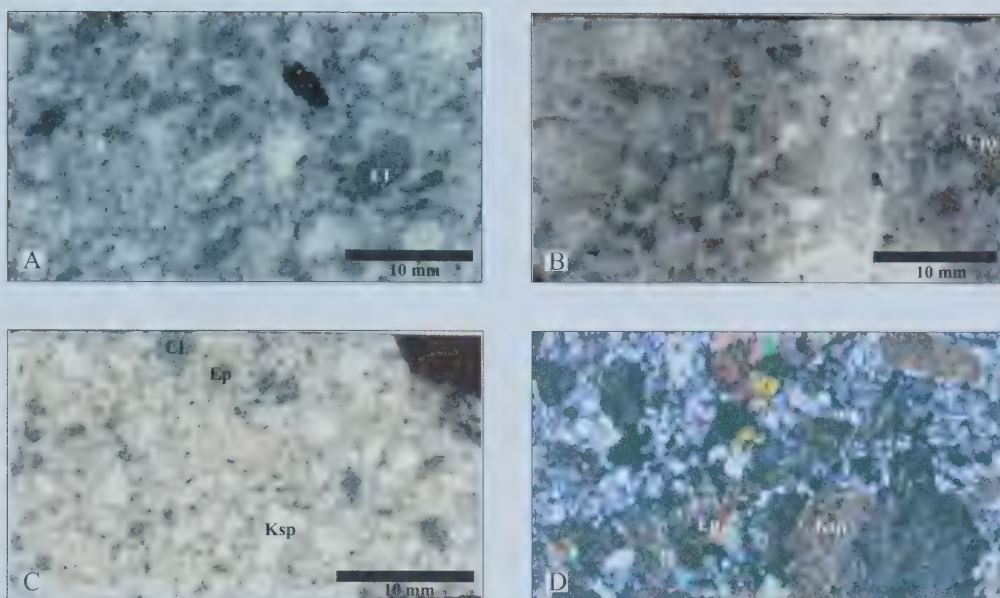


Fig. 5.5. Examples of the early porphyry stock and dikes from the San Carlos and Panantza prospects. A. Hand sample showing relatively fresh early porphyry. Note the crowded texture and lack of quartz eyes. The hornblende has been chloritized (Cl). B. Hand sample from the center of the San Carlos prospect. The hornblende has been replaced by secondary biotite and chalcopyrite (Cpy). Observe the K-feldspar veinlet crosscutting the earlier clay-sericite (green feldspar) alteration in the center right. C. Hand sample from the Panantza prospect of the early porphyry showing alteration of hornblende to chlorite (Cl) and epidote (Ep). The K-feldspar has taken on a bleached appearance indicating alteration to secondary K-feldspar (Ksp). D. Thin section of hand sample C., showing the detail of the alteration. Observe the turbid brown secondary K-feldspar (Ksp) selectively replacing the primary K-feldspar. Epidote (Ep) is replacing hornblende and plagioclase has been altered to sericite. Note the rock maintains much of its primary texture (FOV = 3.35mm, XPL, length of section).

The porphyry has a crowded texture and contains sparse quartz eyes and abundant hornblende phenocrysts. At the Panantza prospect this porphyry forms a dike that intruded into the pink granodiorite phase of the Zamora Batholith. The largest porphyry dike at the Panantza prospect is about 75 x 375m and strikes N to NNE (see map insert). At the San Carlos prospect the porphyry is a small stock with dimensions of about 500 x 300m, which intruded into the white granodiorite of the Zamora Batholith (see map insert).

Late porphyry dikes

A second generation of porphyry dikes intruded both prospects and consists of an intermediate composition, medium-grained, hornblende-plagioclase (granodiorite) porphyry, with accessory bi-pyramidal quartz eyes and euhedral sphene in a groundmass of feldspar/quartz (Fig. 5.6).

At the Panantza prospect the porphyry forms a dike swarm that intruded into the pink granodiorite of the Zamora Batholith and the early porphyry. The porphyry contains large (up to 2cm) phenocrysts of orthoclase, which enclose hornblende and plagioclase (Fig. 5.6D). Two large dikes outcrop at the Panantza prospect: the largest is centered on the mineralization at the prospect and strikes N to NNE, and the second slightly smaller dike strikes W-E and is located some 250m to the ESE of the larger dike (see map insert).

At the San Carlos prospect, three porphyry dikes, striking NNE and spaced about 250m apart, intruded the pink and white granodiorite, the granite, and the early porphyry stock (see map insert). These dikes locally contain abundant volcanic xenoliths closely resembling the dark gray micro-porphyrific andesite previously described from the overlying Oyotún Fm. These xenoliths are interpreted to be the remnants of a carapace breccia much like the one developed nearby between the Zamora Batholith and Oyotún Fm. This relationship is consistent with the interpretation that the level of erosion has only recently exposed the Zamora Batholith in the area.

Post mineral dikes

A suite of intermediate composition hornblende porphyry dikes intruded the San Carlos prospect after the first two stages of mineralization (Fig. 5.7). These dikes strike NE, are thin (<20m wide) and have been altered to a propylitic assemblage (chlorite-

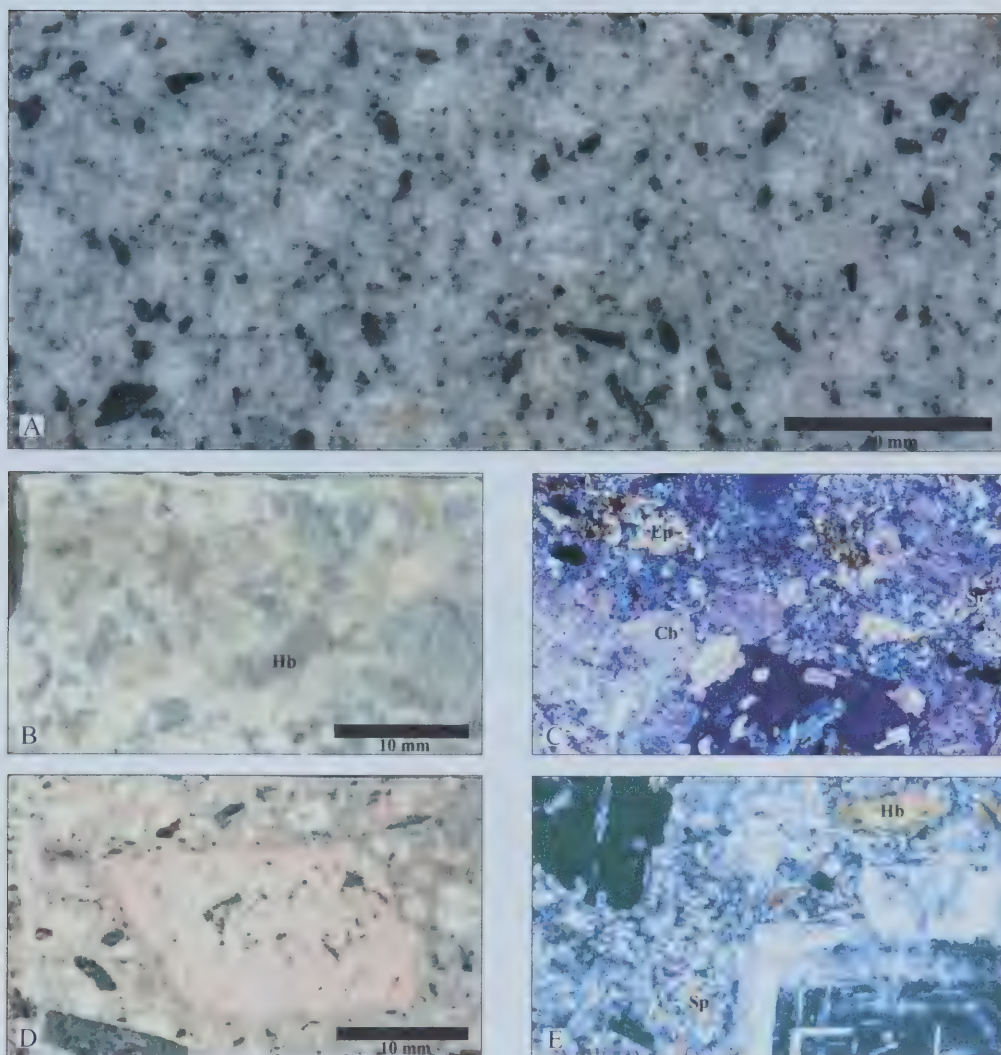


Fig. 5.6. Samples of the late porphyry from the San Carlos and Panatza prospects. Sphene and poikiloblastic K-feldspar crystals are diagnostic of this unit. A. Hand sample from the San Carlos prospect of fresh late porphyry. B. Hand sample from the San Carlos prospect showing moderate phyllic alteration, including replacement of hornblende (Hb) by pyrite. C. Thin section from the San Carlos prospect showing texturally destructive phyllic alteration with an overprint of epidote (Ep)- carbonate (Cb) alteration (propylitic). Note preserved diagnostic euhedral sphene crystal (Sp., FOV = 3.35mm, XPL, length of section). D. Hand sample from the Panatza prospect showing a diagnostic K-feldspar phenocryst (center). E. Thin section from the Panatza prospect showing selective sericite alteration of plagioclase. Note that the hornblende (Hb) is unaltered and the euhedral sphene crystal (Sp., FOV = 3.35mm, XPL, length of section).

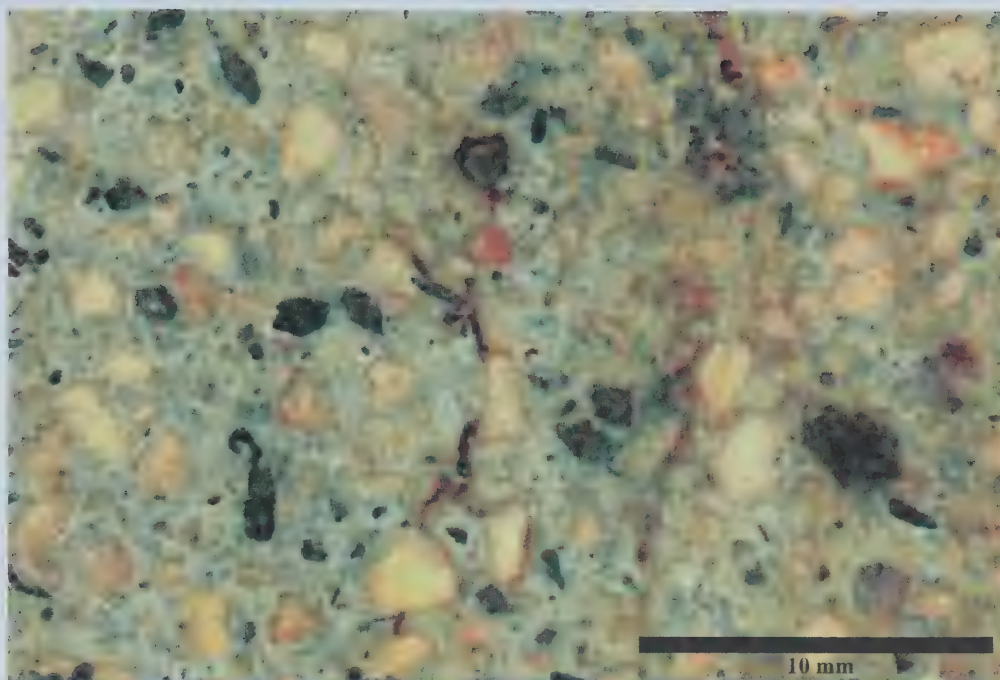


Fig. 5.7. Weathered hand sample from the San Carlos prospect of a post mineral hornblende porphyry dike.

actinolite-calcite) in the NW quadrant of the property but locally to a phyllic assemblage (sericite-pyrite) in the SE quadrant. The change in strike of the porphyry dikes indicates a changing stress field during the life of the hydrothermal system. These dikes are similar to and in roughly the same orientation as the ones observed by Blair (2000) at the Mirador/Wawame prospect.

The youngest intrusive rocks at the prospects consist of a number of small subvolcanic dikes that intruded the older lithologies. The most common type of dike is a dark green micro-porphyritic andesite that in hand specimen is very similar to the previously described andesite of the Oyotún Fm. which it also intrudes. Petrographically the two micro-porphyritic andesites are very different. The Oyotún andesite contains thin euhedral plagioclase laths that are pointed on the terminations and the late andesite contains equant sub to euhedral plagioclase crystals (Fig. 5.8). This unit crops out at both prospects and cross-cuts all lithologies.

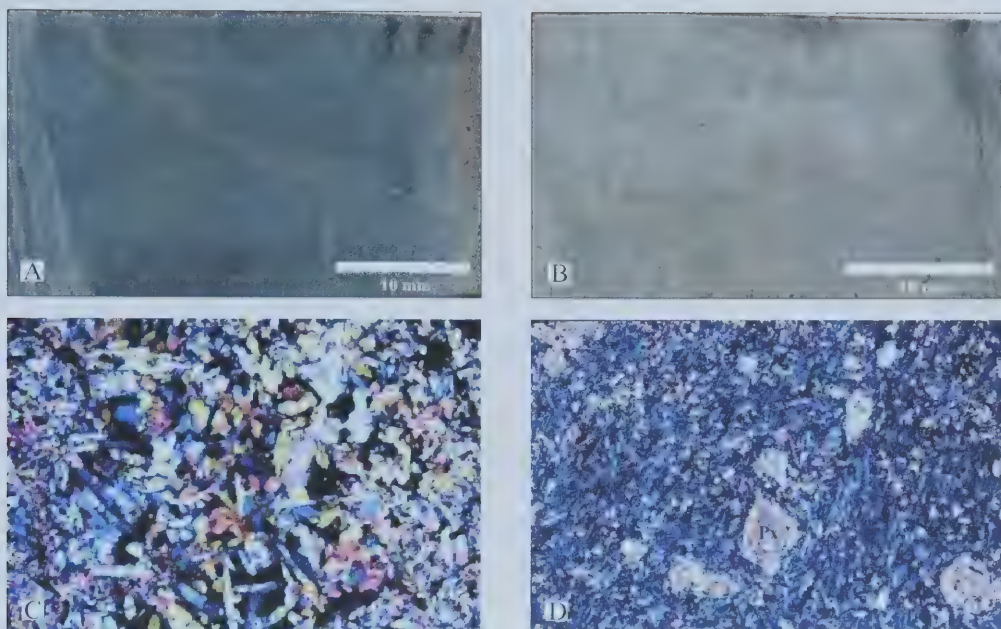


Fig. 5.8. Examples of the post mineral microporphyritic andesite. A. Hand sample of fresh andesite. B. Hand sample of altered andesite. Note the bleached appearance and that magnetite has altered to Fe oxides (reddish spots). C. Thin section from the Panantza prospect showing pervasive actinolite alteration (yellow and purple). Note original magmatic texture has been lost (FOV=3.35mm, XPL, length of section). D. Thin section showing original magmatic texture and carbonate replacement of pyroxene (Px) phenocrysts (FOV = 3.35mm, XPL, length of section).

At the Panantza prospect a subvertical rhyolite dike, up to 10m in width and 1km in length, intrudes both the Oyotún andesite and the pink granodiorite near the contact of the two units. The dike contains local quartz phenocrysts and exhibits flow banding along with devitrification textures. No alteration or mineralization is associated with this dike.

At the San Carlos prospect unmineralized, flesh colored, quartz-eye dacite porphyry dikes similar to the one described above intrudes the white granodiorite (it contains devitrification spherules in thin section). The individual dikes rarely exceed 1m in width and are not altered or mineralized (Fig. 5.9).

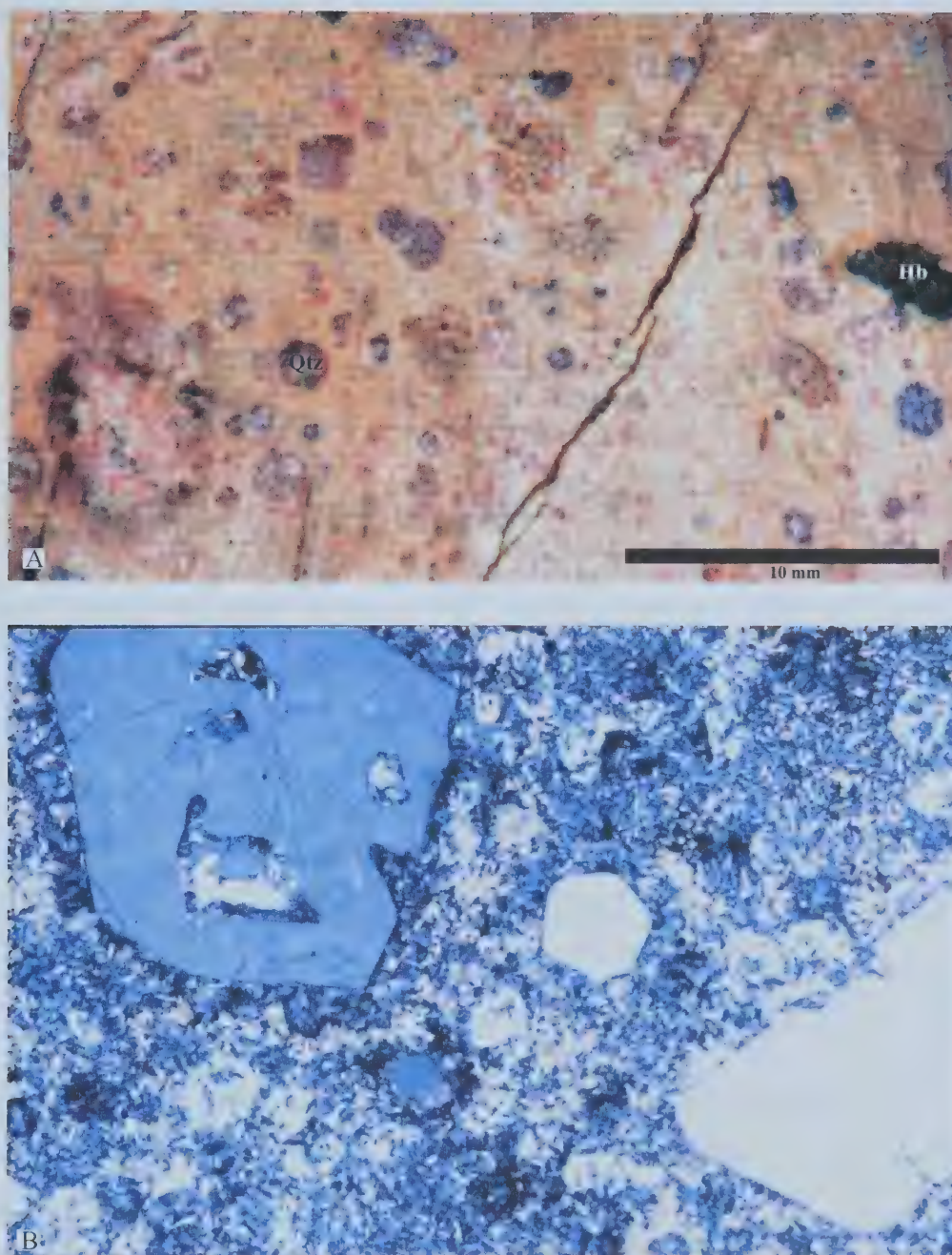


Fig. 5.9. Rhyolite from the San Carlos prospect. A. Hand sample showing abundant quartz eyes (Qtz), rare hornblende (Hb) phenocrysts, and aphanitic matrix. B. Thin section showing quartz eyes and devitrification textures in the matrix (FOV= 3.35mm, XPL, length of section)..

Hornblende andesite porphyry forms a post mineralization pipe/diatreme at the Panantza prospect near the Rio Zamora (see map insert). The andesite intruded the pink granodiorite of the Zamora Batholith and has developed a thin (<15m) contact aureole. No mineralization or alteration has been found associated with this lithology. In the field this unit weathers as large spheroidal boulders.

Hypogene Alteration and Mineralization Characteristics

Radiometric age dating in conjunction with core logging and field mapping has shown that two centers of alteration and mineralization exist at the Panantza prospect and three pulses comprising three centers at the San Carlos prospect. The early centers are located in the NW quadrant of each prospect, whereas the younger centers are found in the SE quadrants. At each prospect the centers overlap to variable degrees (Figs. 5.10 and 5.11).

At the Panantza prospect both centers have been drilled and the older center shows a marked difference in both mineralization and alteration when compared with the younger center. At the San Carlos prospect the older centers have been drilled but not the younger center which is not well exposed (see below).

Panantza

At the Panantza prospect the pink granodiorite of the Zamora Batholith hosts well-developed secondary K-feldspar and magnetite alteration comprising an area of about 1km x 550m (Table. 5.1; Fig. 5.10). Reddish-pink K-feldspar replaces primary plagioclase and the magnetite appears as fine-grained interstitial fillings and fuzzy veinlets. The rock has a black and red color and is strongly magnetic. The mineralization

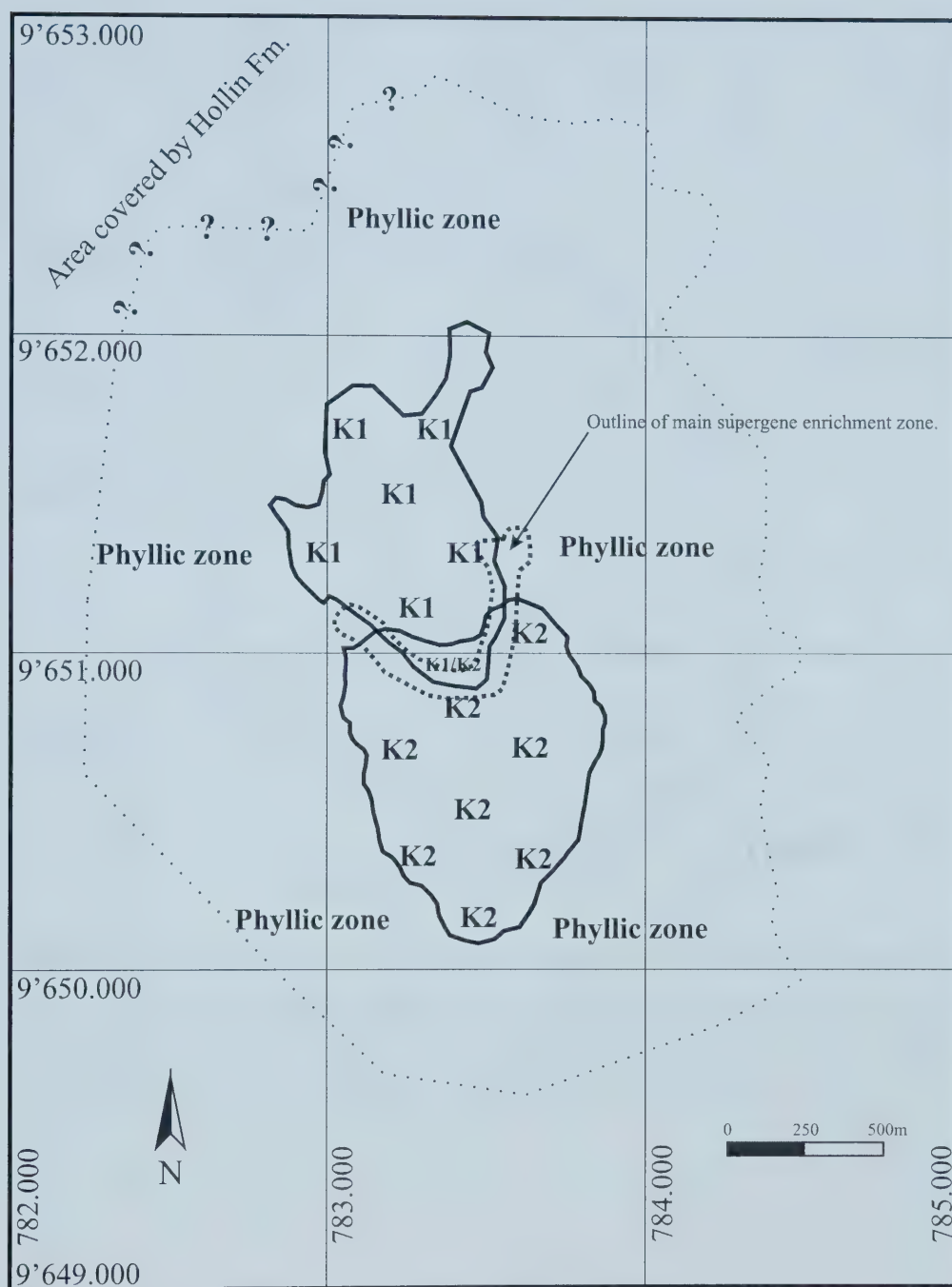


Fig. 5.10. Sketch interpretation map of the Panantza prospect showing the alteration zonation. **K1**= the early K-feldspar and magnetite alteration center. **K2**= the younger K-feldspar and molybdenite center. The two centers of potassic alteration slightly overlap (**K1/K2**) and in turn are surrounded by a phyllic envelope.

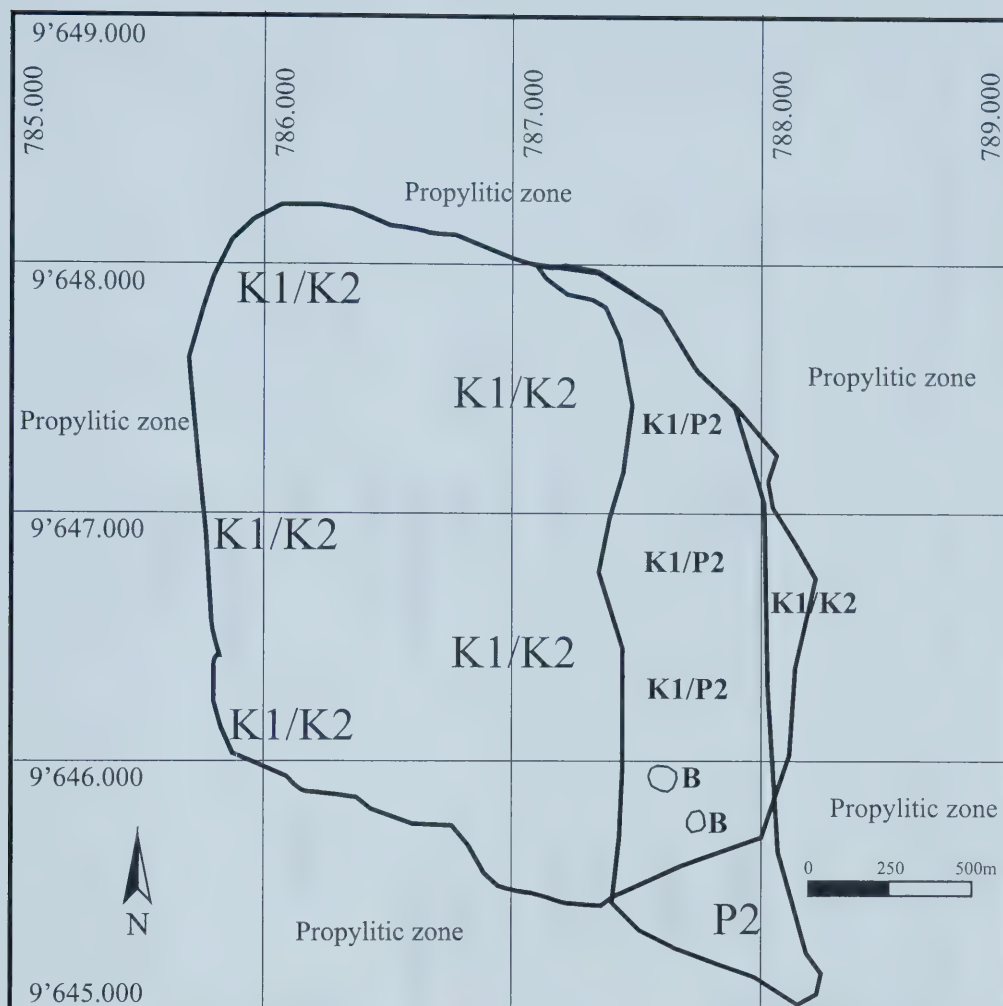


Fig. 5.11. Sketch interpretation map of the San Carlos prospect showing the alteration zonation. **K1/K2**= the early K-feldspar alteration centers. **P2**= the younger phyllic alteration. **B**= matrix supported breccias. **K1/P2**= the phyllic alteration where it overprints the earlier potassic alteration. Both alteration types are surrounded by a propylitic halo.

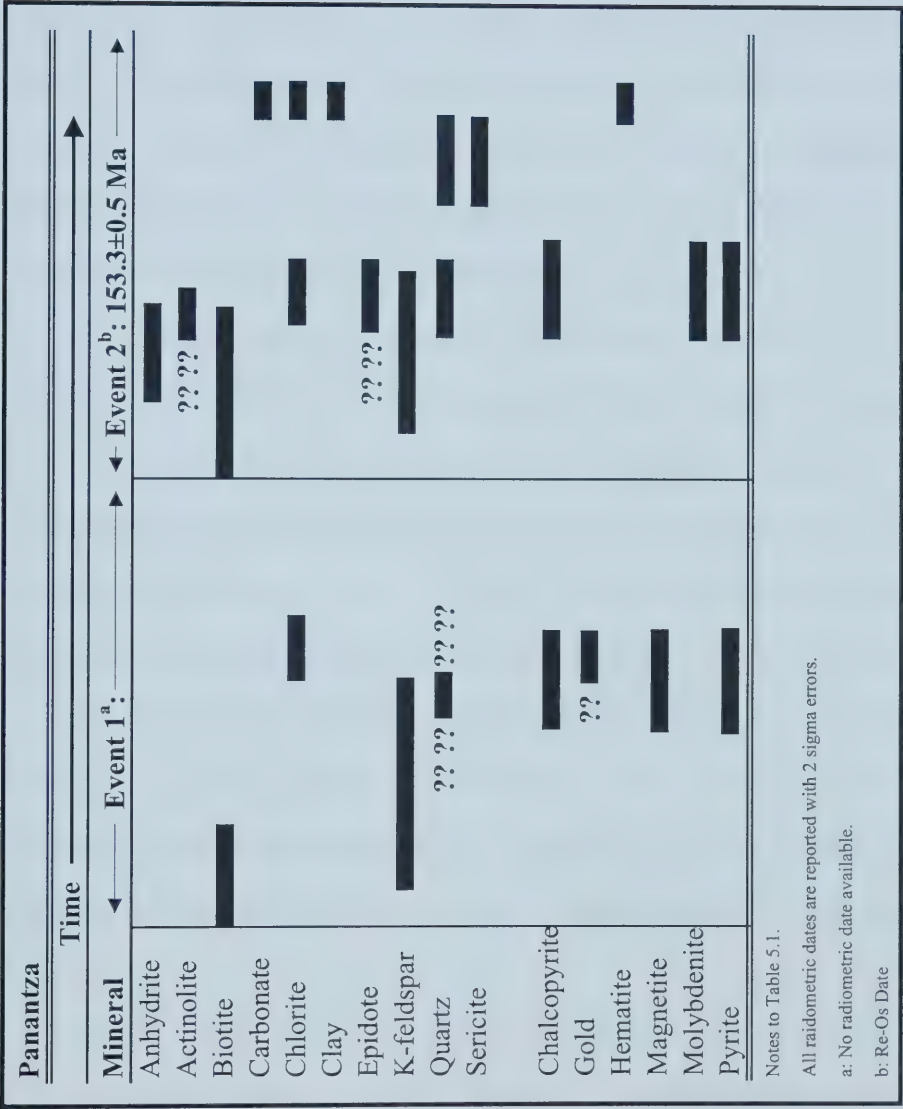


Table 5.1. Paragenetic sequence of hypogene minerals from the Panantza prospect. Observe two events corresponding with different mineral assemblages.

associated with this alteration consists of pyrite and chalcopyrite in a 1 to 1 ratio, with rare bornite, and averages 0.5% copper and 200ppb gold. The host for the gold has not been observed. The early porphyry is exposed at depth in drill core from this area and it is interpreted to be the mineralizer. Intermediate argillic alteration is observed overprinting the potassic alteration. This intermediate argillic overprint affects the plagioclase causing it to turn into greenish clay (probably montmorillonite), whereas mafic minerals such as hornblende are altered into sericite and carbonate.

The second potassic center at the Panantza prospect is located to the SE of the potassic center described above and both centers slightly overlap. This center is also hosted by the pink granodiorite of the Zamora Batholith and is dominated by intense white to slightly pinkish K-feldspar alteration which measures about 900 x 700m in an approximate heart shape (Table. 5.1; Fig. 5.10). The potassic alteration is so intense that the texture of the pink granodiorite has been destroyed and the alteration forms a mappable unit. Both the early and late porphyries are observed in outcrop and drillcore from this area. The early porphyry is strongly altered to an intermediate argillic assemblage, whereas the late porphyry is dominated by potassic alteration containing anhydrite veins and local silicification with a poorly developed intermediate argillic overprint.

Mineralization occurs as disseminations of chalcopyrite with subordinate pyrite and molybdenite (with an average ratio of 5 to 10:1:1 for the sulfides respectively) and only trace amounts of gold. The highest hypogene copper grades at the prospect are observed in this area (>0.75% Cu). No magnetite is associated with this center.

Phyllic alteration (quartz-sericite-pyrite) is pervasively developed along the eastern margin of the prospect (Fig. 5.10). Along the south and west margins of the prospect phyllic alteration is only moderately developed, but the northern margin of the prospect was not observed. All lithologies including the Oyotún andesite, pink granodiorite, and early and late porphyry dikes have been affected to some extent by this alteration type. The late porphyry dike that strikes E-W along the eastern edge of the prospect can be traced into the pervasively developed phyllic zone until the only way to identify the dike is to pick out the remnant quartz eyes from the sericite. The Oyotún andesite shows only very localized development of the phyllic alteration which consists of veinlets that appear to follow joints. These veinlets have a bleached selvage and are quite easy to recognize in the field (Fig. 5.12). The pink granodiorite of the Zamora Batholith is altered to a phyllic assemblage including K-feldspar potassic alteration overprinted by phyllic alteration.

Propylitic alteration (epidote-chlorite-calcite) is weakly developed at the prospect and is only rarely observed. It forms in the more mafic rocks such as the Oyotún andesite. Rare sphalerite and galena veins are found within this alteration type along the outer margins of the prospect (Fig. 5.13).

Intermediate argillic alteration (sericite-carbonate-clay) affects all lithologies in a variable manner at the prospect. Some of this alteration appears to be related to supergene processes and will be discussed below.

San Carlos

The older potassic center at the San Carlos prospect is hosted within the early porphyry stock and the white granodiorite of the Zamora Batholith (Fig. 5.11). The first



Fig. 5.12. Outcrop of Oyoútún Fm. andesite from the Rio Panantza at the Panantza prospect hosting veins and veinlets of phyllic alteration (tan).

alteration event is dominated by secondary biotite and epidote replacing primary biotite and hornblende. Sparse K-feldspar veinlets containing sulfide minerals also cut the area of secondary biotite alteration. A retrograde intermediate argillic (sericite-chlorite-carbonate-clay-hematite) overprints the potassic alteration. A second potassic alteration event overprints the first event and is comprised of a slightly different assemblage: secondary biotite-K-feldspar-anhydrite-epidote (Table. 5.2; Fig. 5.14). This potassic assemblage overprints the alteration from the first event. A retrograde intermediate argillic overprint also developed at the end of this event. The mineralization associated with these two events consists primarily of disseminations, veinlets, and replacement of mafic minerals by chalcopyrite and pyrite. Molybdenite has been observed in quartz veins with K-feldspar and sericite selvages, and as rare disseminations.

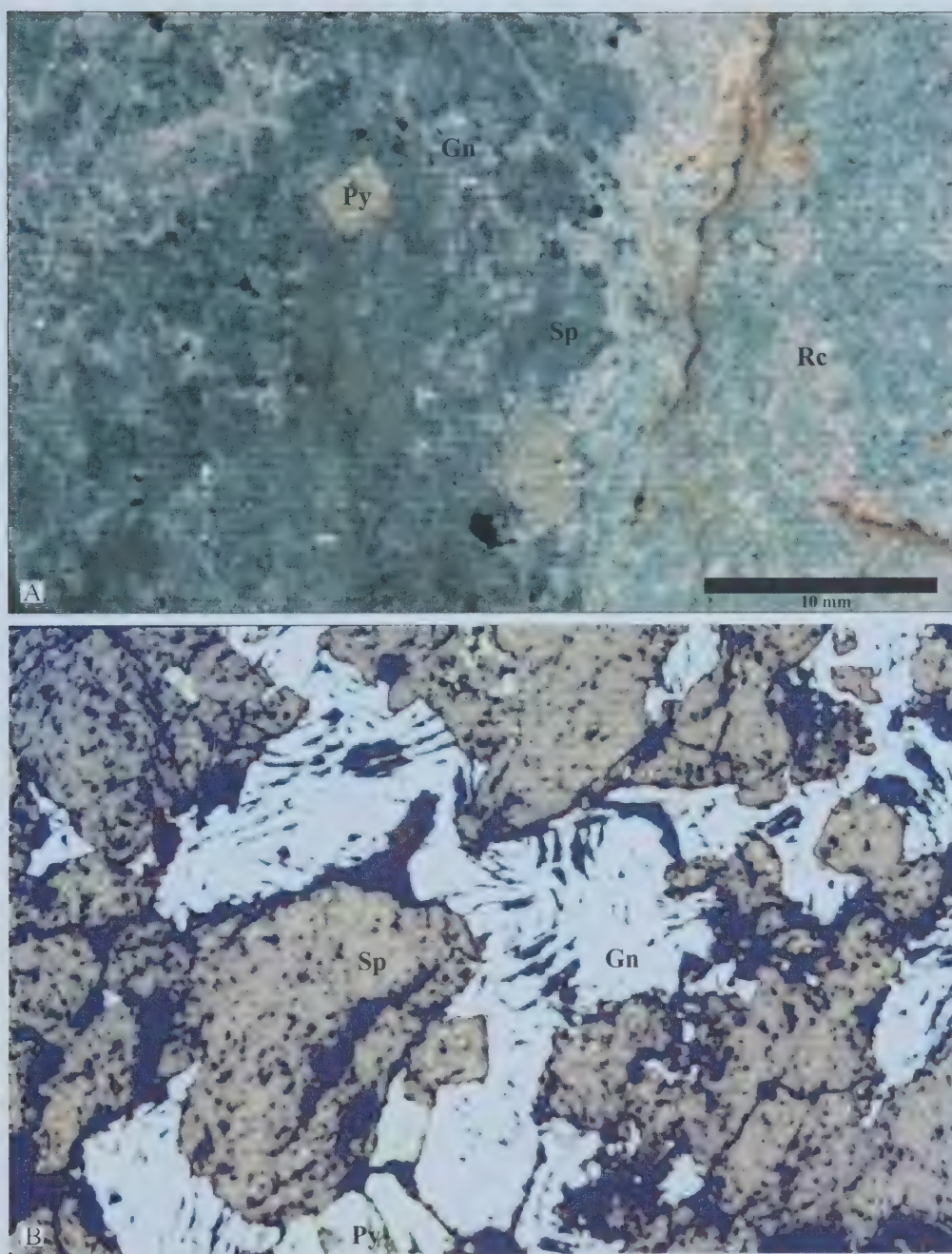


Fig. 5.13. Example of Pb-Zn-Cu vein from the periphery of the Panantza prospect. A. Hand sample showing contact of the vein with the pink Zamora Batholith granodiorite. Observe the pink alteration mineral rhodochrosite (Rc) and the massive texture of the vein. Galena (Gn), pyrite (Py), and sphalerite (Sp). B. Thin section showing detail of the relationship between sulfide phases Galena (Gn), pyrite (Py), and sphalerite (Sp) with chalcopyrite disease (FOV=3.35mm, RL, length of section).

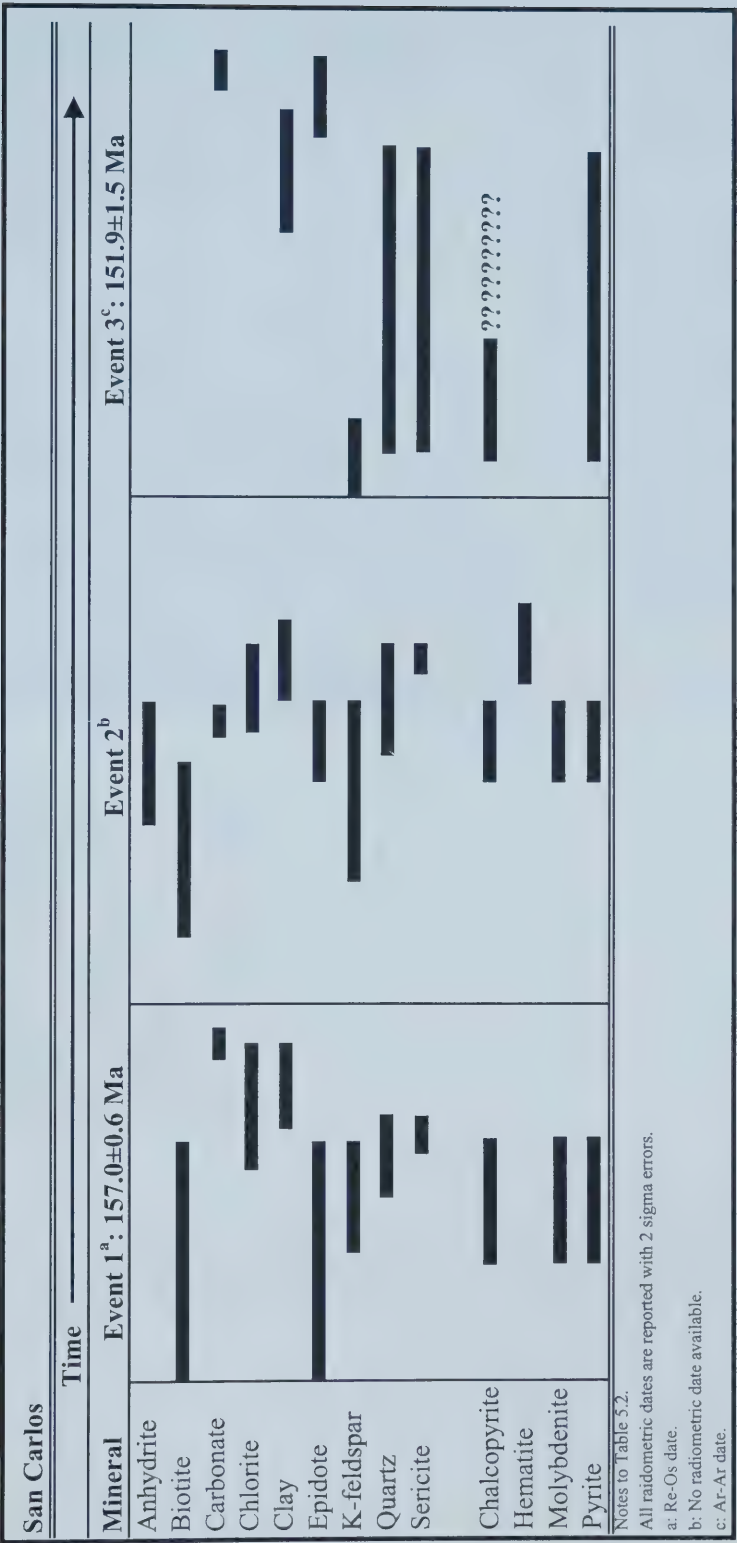


Table 5.2 Paragenetic sequence of hypogen minerals at the San Carlos prospect. Three mineralization events occurred, two of which have been radiometrically dated. The middle event is observed in drill core.

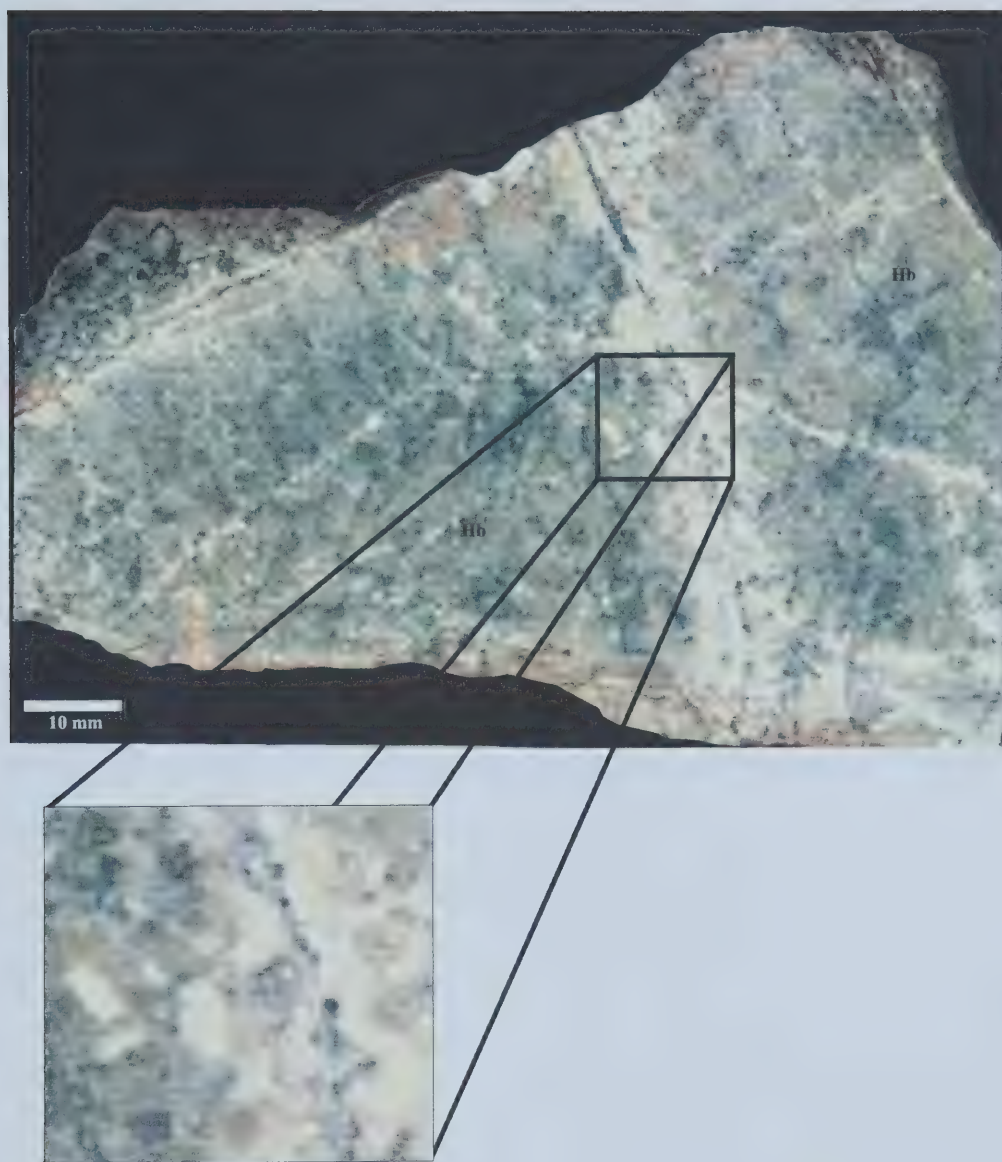


Fig. 5.14. Hand sample of strongly altered early porphyry from the San Carlos prospect showing multiple sets of K-feldspar veinlets (white and pink) overprinting intermediate argillic alteration (green). Primary hornblende (Hb) has been replaced by chalcopyrite and secondary biotite, whereas primary feldspars have been altered to sericite and clay (green).

This alteration center is different than the early K-feldspar and magnetite center at the Panantza prospect because it does not contain abundant magnetite or gold in the >150 ppb range. The early porphyry stock hosts the highest copper grades, although the white granodiorite locally hosts equivalent grades (>0.7% copper).

Field mapping, drill core logging, PIMA data, and radiometric age dating have provided evidence for a third alteration center. The center is not well exposed at surface and has only been drilled along its fringes (DDHs 1,2,10,11, and 23). Drill core from the fringes of the center contain veins and veinlets with increasing K-feldspar and coarse-grained muscovite-selvages with depth in the late porphyry (DDH 23).

Two quartz and muscovite matrix-supported breccias have been mapped in this area near the contact between the granite and Zamora Batholith granodiorite. The alteration within these breccias consists of coarse-grained muscovite supporting sericite-altered clasts of the granite. The coarse-grained muscovite centers are in turn enveloped by fine-grained sericite and illite halos, which locally pervasively alter the easternmost late porphyry dike, Zamora Batholith granodiorite, and granite (Table. 5.1; Figs. 5.11 and 5.15). Coarse-grained muscovite-selvages quartz and sulfide veins along with rare K-feldspar veins appear to radiate out from these centers and are moderately well developed at depth within the late porphyry (e.g. DDH 23). This type of vein assemblage is consistent with a potassic alteration assemblage (Thompson and Thompson, 1996).

The breccias are spatially associated with pyrite, chalcopyrite, and molybdenite. The pyrite and chalcopyrite occur as disseminations within the coarse-grained muscovite, locally attaining 3% of the rock. Pyrite and trace chalcopyrite replaces hornblende in the altered late porphyry dike, whereas the molybdenite occurs as quartz-molybdenite veins

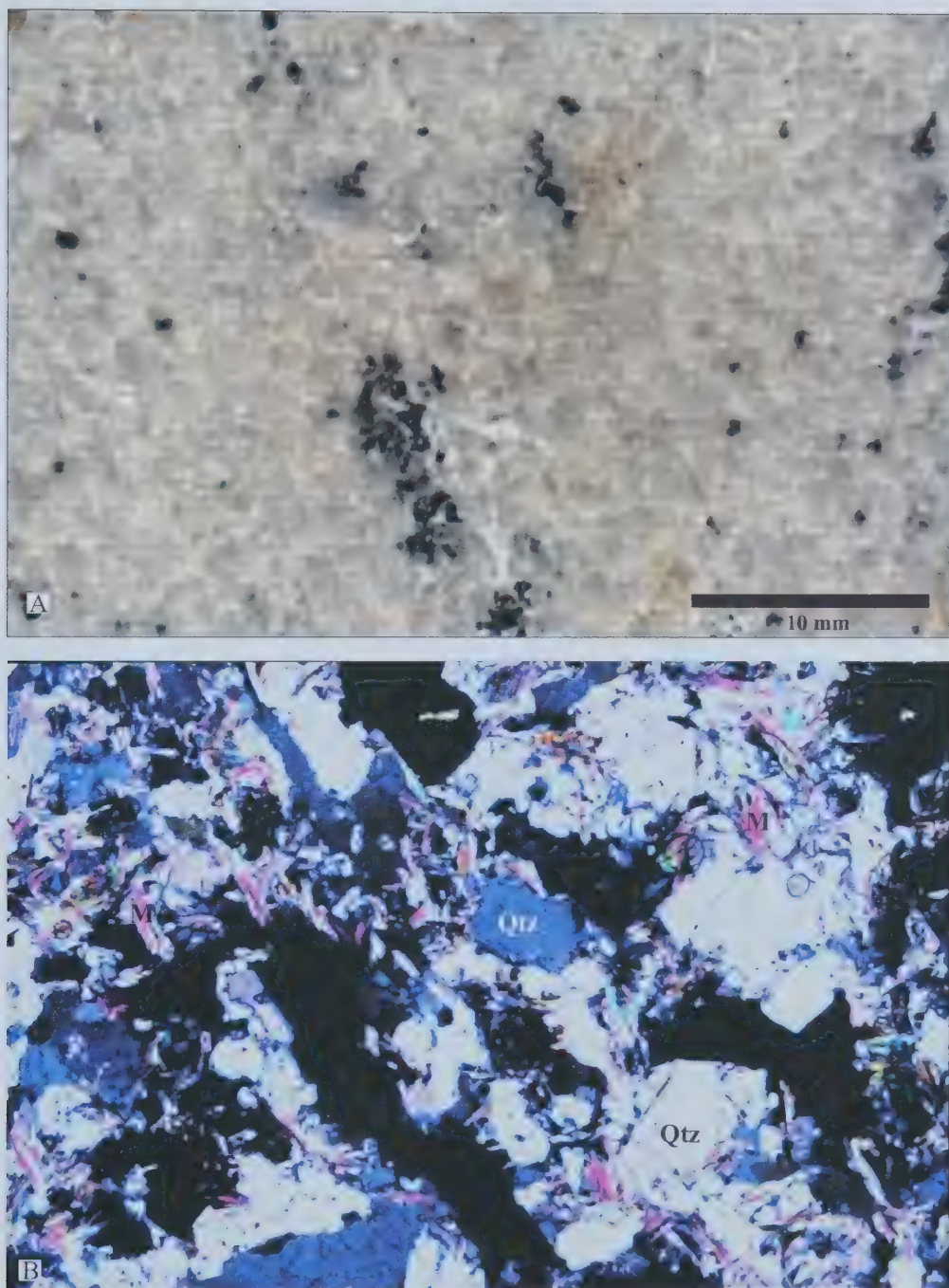


Fig. 5.15. Examples of coarse-grained muscovite and quartz alteration from the younger center at the San Carlos prospect. A. Hand sample showing complete destruction of the original igneous texture. B. Thin section showing abundant muscovite (M) and quartz (Qtz) along with sulfides (FOV = 3.35mm, XPL, length of section).

around the periphery of the breccias. At this point it is unknown if the younger center contains mineralization of the same tenor as the later mineralization at the Panantza prospect; however, surface grades locally attain 0.5% copper as disseminations along with subordinate molybdenite veinlets. As at the Panantza prospect, this center appears to be slightly offset to the SE from the older center.

Propylitic alteration is moderately to well-developed along the SW margin of the prospect where it overprints potassic alteration. This alteration assemblage consists of epidote-pyrite-carbonate. Some propylitic alteration occurs within the region where the two centers overlap but it is only poorly developed and is interpreted as an outer alteration halo to the third center.

Within the Oyotún Fm. near the contact of the white granodiorite, some sparsely developed skarn minerals have been observed (mainly garnets, but some pyrrhotite is also observed). This alteration type is not readily apparent in the field but is easily observed in thin section. No copper mineralization is associated with this alteration type in the San Carlos area, although further to the north at the Kutucus prospect it is well developed (Sillitoe, 1998).

Supergene Oxidation and Enrichment

Both prospects show some local development of a secondary chalcocite enrichment blanket. The lack of abundant pyrite associated with the porphyries dictates in-situ oxide development which, is moderately developed at both prospects (Chávez, 2000; Gendall, 2000). The grade of the oxide zone is generally consistent with the hypogene grades although some enrichment has occurred. There is an upper leached zone dominated by Fe-Mn oxides and clay. Below this Fe-Mn oxide zone is a zone of clay and

copper oxides that give the rock a green-blue color which in turn is followed by a zone containing locally developed secondary sulfide minerals (chalcocite and covellite).

The oxides are better developed at the Panantza prospect where they form a 500 x 300m blanket of variable thickness centered on the overlap between the two potassic alteration centers (Fig. 5.10). The source of the oxides appears to be the early potassic center, which contains a higher pyrite to chalcopyrite ratio and is situated up slope from the younger potassic center. With the abundant rainfall in the region, oxidation of the sulfides occurs quite readily and down slope transport of acidic copper enriched water occurs. When this water comes into contact with the chalcopyrite rich younger potassic center the acid attacks the feldspars and mafic minerals altering them to clay and carbonate minerals.

6. Whole-Rock Lithogeochemistry of Igneous Rocks from the Pangui District

Major and trace element methodology

Twenty-seven samples of least altered igneous rocks from the San Carlos and Panantza prospects have been analyzed for major and trace elements using X-ray fluorescence (XRF) and inductively-coupled plasma mass spectrometry (ICP-MS) techniques. The samples were prepared by crushing and milling in an agate disc mill, and 2g splits of this material were fused with Li-tetraborate to make glass pellets for XRF analysis at XRAL Laboratories (Don Mills, Ontario, Canada). Further aliquots of powder were analyzed by ICP-MS for trace elements at the University of Saskatchewan (method of Jenner et al., 1990).

Replicate analyses of international standards indicate accuracy within 10 relative percent of the standard values for the reported trace elements analyzed by ICP-MS, and to within 1 and 5 relative percent for major and minor elements respectively analyzed by XRF.

The samples were selected for their relative lack of alteration, but all of the samples have been altered to a minor degree. It is probable that some elements, particularly alkalis and silica, have been added or lost. Zirconium values may also be artificially low due to the possibility of incomplete acid digestion of zircon in the laboratory. Where possible the data have been plotted with data from a study of Jurassic rocks from northern Perú and Ecuador by Romeuf et al. (1995).

Major elements

Major element analyses for the San Carlos and Panantza prospects are presented in Tables 6.1. and 6.2. respectively. The data were plotted on a $\text{FeO}^* - \text{MgO} - (\text{Na}_2\text{O} +$

Table 6.1. Major (XRF) and Trace element (ICP-MS) data for igneous rock samples from the San Carlos prospect

Sample no.	56	44	S3	118A	85	SC1102	SC0201	95B	105	113A	138	SC0806	SC0315	SC0813	126B
	PGDR	Granite	Granite	WGDR	WGDR	WGDR	EP	LP	LP	LP	LP	LP	LP	LP	Rhyolite
MAJOR ELEMENTS (wt. %)															
SiO ₂	68.1	78.2	78.4	65.7	63.9	67.1	64.2	62.8	66.6	62.6	65.1	64.5	68.9	64.9	72.9
Al ₂ O ₃	14.6	12.4	12.0	15.0	15.3	14.9	16.7	17.7	15.4	17.1	16.3	16.1	15.6	16.4	15.7
CaO	2.9	0.2	0.2	2.6	2.7	3.8	5.1	6.0	3.8	3.7	5.2	4.6	3.4	5.3	0.0
MgO	1.1	0.0	0.0	1.4	1.6	1.5	1.2	1.3	1.2	1.3	1.1	1.2	1.1	1.2	0.1
Na ₂ O	3.1	3.8	2.9	3.0	2.4	2.2	2.7	3.4	2.6	4.9	3.3	3.1	2.7	3.3	0.1
K ₂ O	4.2	3.8	4.4	4.0	4.9	3.1	1.9	1.8	3.2	2.3	2.7	3.6	3.6	2.7	5.4
Fe ₂ O ₃	2.9	0.6	0.9	4.2	4.7	4.6	5.2	4.8	4.5	5.0	4.2	3.7	2.9	4.0	1.4
MnO	0.1	0.0	0.0	0.1	0.0	0.0	0.1	0.1	0.1	0.2	0.1	0.1	0.1	0.1	0.0
TiO ₂	0.3	0.1	0.2	0.3	0.4	0.3	0.4	0.4	0.3	0.4	0.3	0.3	0.3	0.3	0.1
P ₂ O ₅	0.1	0.0	0.0	0.1	0.2	0.1	0.2	0.2	0.1	0.2	0.2	0.2	0.1	0.2	0.0
Cr ₂ O ₃	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
LOI	1.1	1.1	1.1	2.9	2.5	2.0	2.6	1.8	2.0	1.7	0.8	1.2	1.3	0.7	3.5
Total	98.5	100.1	100.1	99.2	98.5	99.7	100.1	100.2	99.7	99.4	99.2	98.6	100.0	98.9	99.2
TRACE ELEMENTS (ppm)															
Cs	1.3	3.6	2.7	2.6	1.6	2.4	1.4	0.7	1.4	0.7	1.2	1.1	1.0	0.9	2.4
Li	6	2	1	9	10	5	10	4	5	13	2	5	10	4	16
Rb	101	134	109	90	105	85	46	44	74	55	61	53	74	47	188
Sr	474	59	140	420	481	523	585	663	540	599	563	608	505	545	72
Ba	985	85	151	1123	1295	1075	684	1093	1262	841	1137	1159	1464	1131	1263
Nb	4.7	4.9	4.7	4.4	3.6	4.8	1.5	4.1	4.3	4.1	4.2	3.7	3.8	4.1	5.2
Ta	0.5	0.5	0.4	0.4	0.3	0.5	0.1	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.4
Zr	20	76	27	22	14	19	13	34	19	24	15	28	28	28	69
Y	11	4	3	12	13	11	14	17	13	17	15	14	12	14	7
La	18	21	15	17	19	17	16	17	21	16	15	15	19	16	20
Ce	30	30	20	32	34	31	31	34	38	31	28	28	34	31	27
Pr	3.3	2.4	1.6	3.6	4.0	3.4	3.7	4.3	4.1	3.9	3.5	3.4	3.7	3.6	3.0
Nd	12	6	5	14	15	13	15	18	16	16	14	13	14	15	10
Sm	2.3	0.8	0.6	2.6	3.0	2.6	3.1	3.7	3.0	3.4	3.0	2.8	2.7	3.1	1.4
Eu	0.6	0.1	0.2	0.7	0.8	0.7	0.8	1.1	0.8	1.0	0.9	0.8	0.8	1.0	0.4
Gd	2.0	0.7	0.5	2.3	2.8	2.3	2.9	3.6	2.6	3.4	2.8	2.7	2.4	2.9	1.1
Tb	0.28	0.08	0.06	0.32	0.36	0.31	0.40	0.50	0.34	0.48	0.37	0.38	0.29	0.39	0.15
Dy	1.8	0.6	0.4	2.0	2.2	2.0	2.4	3.1	2.2	3.1	2.5	2.4	1.9	2.4	1.1
Ho	0.37	0.12	0.08	0.42	0.45	0.40	0.51	0.64	0.46	0.62	0.52	0.48	0.38	0.52	0.24
Er	1.14	0.48	0.27	1.21	1.27	1.16	1.44	1.87	1.30	1.79	1.51	1.39	1.17	1.54	0.79
Tm	0.18	0.10	0.05	0.20	0.20	0.18	0.23	0.29	0.21	0.28	0.24	0.24	0.17	0.24	0.14
Yb	1.35	0.95	0.51	1.45	1.42	1.32	1.55	1.96	1.46	1.88	1.68	1.67	1.19	1.69	1.18
Lu	0.22	0.19	0.09	0.23	0.23	0.21	0.23	0.29	0.24	0.28	0.26	0.26	0.21	0.25	0.22
Hf	1.9	3.2	2.0	2.1	1.0	2.0	0.7	1.4	0.9	1.5	0.8	1.3	1.9	1.8	2.8
Pb	11	8	9	12	8	6	5	4	8	11	7	5	6	6	9
Bi	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Th	9	26	9	9	16	4	3	3	7	3	4	4	6	4	8
U	2.6	5.7	2.0	2.9	3.5	1.8	0.8	1.3	2.0	1.2	1.5	1.4	1.4	1.3	3.0
Sc	6	0	0	9	9	9	7	7	7	8	7	6	6	7	1
V	67	5	5	93	100	104	65	83	80	88	65	66	57	64	14
S	0.4	0.2	0.5	2.3	2.0	1.7	2.2	1.4	1.3	0.0	0.2	0.4	0.5	0.2	0.0
CO ₂	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Eu ₀ /Eu _T	0.9	0.5	1.1	0.9	0.9	0.9	0.8	0.9	0.8	0.9	1.0	0.9	1.0	1.0	0.9
La _T /Yb _T	9.4	15.9	20.9	8.6	9.4	9.2	7.3	6.1	10.4	6.1	6.3	6.3	11.3	6.8	12.2
La _T /Sm _T	4.9	16.6	16.2	4.2	4.0	4.2	3.3	3.0	4.5	3.0	3.2	3.5	4.5	3.4	9.3

Notes:

Accuracy is estimated to be within 10 relative percent for trace elements, and to within 1 relative percent for major elements (± 0.1 wt. % at levels below 2 wt. %).Eu₀/Eu_T = Eu₀ x Gd_TLa_T/Yb_T: C1 normalization values of Sun and McDonough (1989)La_T/Sm_T: C1 normalization values of Sun and McDonough (1989)

Major elements analyzed by XRF at XRAL Laboratories, Don Mills, Ontario, Canada

Trace elements analyzed by ICP-MS at the University of Saskatchewan, Department of Geological Sciences, analyst Qianli Xie

PGDR = Pink Zamora Batholith, WGDR = White Zamora Batholith, EP = Early porphyry, LP = Late porphyry

Table 6.2. Major (XRF) and Trace element (ICP-MS) data for igneous rock samples from the Panamint prospect

Sample no	P55	P43	PA0603	PA0204	P9	P12	P18	P59	PA0408	P60	P58	P15
	PGDR	PGDR	PGDR	EP	EP	LP	LP	Rhyolite	Rhyolite	Oyotun Fm.	Oyotun Fm.	Andesite
MAJOR ELEMENTS (wt. %)												
SiO ₂	68.3	61.8	70.5	65.7	61.2	67.1	65.8	75.9	76.3	60.7	59.3	57.6
Al ₂ O ₃	15.5	15.9	15.2	15.9	16.9	16.4	15.8	13.6	14.7	18.0	16.8	17.1
CaO	0.4	3.3	1.4	2.1	4.9	1.1	4.1	0.1	0.1	5.9	4.5	4.7
MgO	1.6	1.9	1.5	1.4	1.4	1.1	1.0	0.1	0.1	2.2	3.1	3.3
Na ₂ O	4.2	2.5	2.6	3.1	3.7	5.1	3.3	3.3	0.1	4.4	2.6	2.9
K ₂ O	3.6	4.0	4.0	4.6	2.9	2.8	3.6	4.4	4.1	1.7	3.0	2.0
Fe ₂ O ₃	3.8	5.3	2.8	3.0	5.3	4.2	3.6	1.2	0.7	5.3	7.8	7.2
MnO	0.1	0.5	0.1	0.1	0.3	0.2	0.3	0.0	0.1	0.1	0.4	0.3
TiO ₂	0.3	0.4	0.3	0.3	0.4	0.3	0.3	0.2	0.0	0.6	0.7	0.7
P ₂ O ₅	0.1	0.2	0.1	0.2	0.2	0.1	0.2	0.0	0.0	0.3	0.2	0.2
Cl ₂ O ₃	0.0	0.0	0.0	0.0	0.0	0.1	0.0	0.0	0.0	0.0	0.0	0.0
LOI	2.1	2.8	1.5	2.3	1.7	1.6	2.2	1.2	3.2	0.9	1.7	2.9
Total	100.0	98.6	99.9	98.6	98.9	100.1	100.0	99.9	99.4	100.0	100.1	98.9
TRACE ELEMENTS (ppm)												
Ca	4.4	4.1	1.5	4.8	1.3	1.2	2.2	3.1	8.3	0.5	5.6	3.9
Li	12	14	19	16	9	10	17	3	59	11	15	22
Rb	95	96	94	106	53	68	95	104	144	36	81	51
Sr	214	609	435	299	618	246	580	107	357	1038	515	501
Ba	849	870	1189	1417	1475	786	1113	1656	991	1143	1148	1043
Nb	3.2	5.2	5.4	3.6	4.5	4.2	3.5	6.5	17.7	3.8	6.5	6.2
Ta	0.2	0.4	0.5	0.2	0.3	0.3	0.2	0.6	1.3	0.2	0.4	0.4
Zr	13	9	14	28	27	38	35	92	25	77	41	74
Y	18	21	13	9	17	14	12	10	18	14	26	23
La	21	19	16	11	18	19	16	15	14	20	20	19
Ce	38	36	30	21	36	31	28	23	28	33	40	39
Pr	4.7	4.7	3.6	2.5	4.5	3.9	3.1	2.7	3.5	5.2	5.1	4.8
Nd	18	19	14	10	18	15	12	9	13	22	22	20
Sm	3.6	4.0	2.9	2.1	3.5	3.1	2.4	1.8	3.4	4.1	5.0	4.5
Eu	0.9	1.0	0.7	0.7	1.1	0.9	0.7	0.4	0.5	1.2	1.1	1.1
Gd	3.4	3.8	2.6	1.8	3.4	2.7	2.3	1.6	3.1	3.4	4.9	4.6
Th	0.50	0.55	0.37	0.26	0.48	0.38	0.30	0.24	0.51	0.43	0.77	0.65
Dy	3.0	3.5	2.3	1.6	2.9	2.4	1.9	1.6	3.0	2.4	4.8	4.4
Ho	0.63	0.73	0.48	0.32	0.60	0.51	0.41	0.32	0.57	0.44	0.98	0.88
Er	1.88	2.16	1.44	1.01	1.80	1.53	1.23	1.00	1.53	1.15	2.82	2.43
Tm	0.29	0.33	0.22	0.14	0.28	0.22	0.19	0.18	0.22	0.17	0.43	0.35
Yb	2.06	2.08	1.65	1.05	1.80	1.66	1.42	1.25	1.43	1.02	2.71	2.40
Lu	0.31	0.31	0.24	0.17	0.29	0.27	0.22	0.21	0.18	0.14	0.39	0.34
Hf	1.5	0.7	1.7	2.0	1.3	2.6	1.7	2.9	1.6	2.7	1.4	3.3
Pb	6	49	10	11	16	5	16	7	13	13	15	18
Bi	0	2	0	0	0	0	0	0	0	0	0	0
Tl	9	5	12	5	5	7	5	11	7	2	6	7
U	2	3.7	3.1	1.6	1.6	2.6	2.1	3.3	3.9	0.7	1.5	1.2
Sc	8	11	8	6	8	6	6	2	2	12	23	25
V	87	120	77	84	99	73	68	7	2	131	176	172
S	0.73	1.75	0.33	1.22	0	0	0	0	0	-0.01	0.13	0
CO ₂	0	0	0	0	0	0	0	0	0	0.01	0.03	0
Eu _{0.5} Eu _{0.5}	0.82	0.74	0.83	1.07	0.97	0.93	0.91	0.71	0.50	1.01	0.66	0.74
La _{0.5} Yb _{0.5}	7.19	6.61	6.97	7.69	7.10	8.37	7.94	8.40	7.14	14.22	5.25	5.65
La _{0.5} Sm _{0.5}	3.75	3.12	3.61	3.46	3.36	4.04	4.23	5.28	2.73	3.15	2.58	2.71

Notes:

Accuracy is estimated to be within 10 relative percent for trace elements, and to within 1 relative percent for major elements (± 0.1 wt. % at levels below 2 wt. %).Eu_{0.5}Eu_{0.5}, Eu_{0.5}, Sm_{0.5}, & Gd_{0.5}.La_{0.5}Yb_{0.5}, Cl normalization values of Sun and McDonough (1989).La_{0.5}Sm_{0.5}, Cl normalization values of Sun and McDonough (1989).

Major elements analyzed by XRF at XRAL Laboratories, Don Mills, Ontario, Canada.

Trace elements analyzed by ICP-MS at the University of Saskatchewan, Department of Geological Sciences, analyst Qianli Xie.

PGDR = Pink Zamora Batholith; EP = Early porphyry; LP = Late porphyry.

K₂O) ternary diagram (after Irving and Baragar, 1971; Fig. 6.1a) and show a calc-alkaline trend for the sample suite which, supports the calc-alkaline trend observed by Zamora and Litherland (1993a) for the Zamora Batholith. The intrusive and volcanic rocks generally plot within the subalkaline field on a TAS diagram overlain with sigma isopleths (after Rittmann, 1957; Fig. 6.1b). Plotting the data on a TiO₂ vs. Al₂O₃ diagram (after Irving and Baragar 1971; Fig. 6.2a) shows that the data from this study and Romeuf et al. (1995) fall into the arc related field.

The volcanic samples were classified according to the IUGS scheme of LeMaitre et al. (1989) and are presented in Fig. 6.1b. Volcanic rocks from both prospects plot within the andesite and rhyolite fields. Samples P58 and P60 are from the Oyoútún Fm. and sample P15 is an andesite that may be Early Cretaceous, possibly belonging to the Misahualli Member of the Chapiza Fm. The rhyolites from both prospects are believed to be the same approximate age as the andesite P15. This would indicate that the Misahualli Member hosts a bimodal volcanic suite, an observation consistent with a description of bimodal volcanic rocks within the Misahualli Member by Hall and Calle (1982).

The intrusive samples were classified using the IUGS QAP modal classification scheme of Streckeisen (1976) and are presented in Fig. 6.2b. The majority of the samples plot within the high-K field of the SiO₂ vs. K₂O diagram of LeMaitre et al. (1989; Fig. 6.3) and all of the seven samples that plot outside of the high-K field plot within the medium-K field of the diagram.

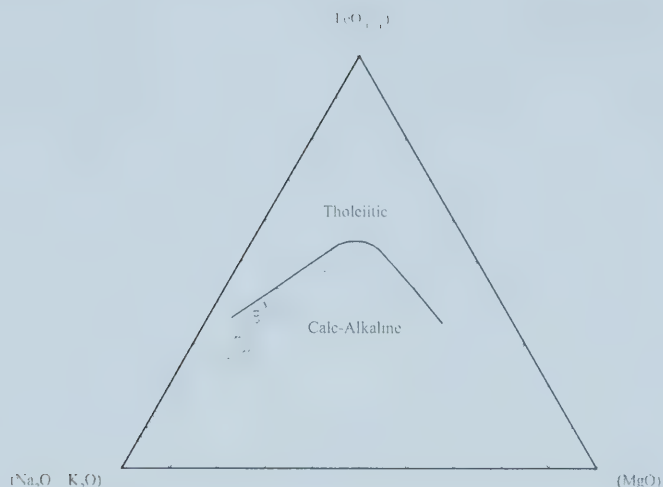


Fig.6.1a. Data plotted on a ternary diagram after Irvine and Baragar (1971) showing the calc-alkaline trend.

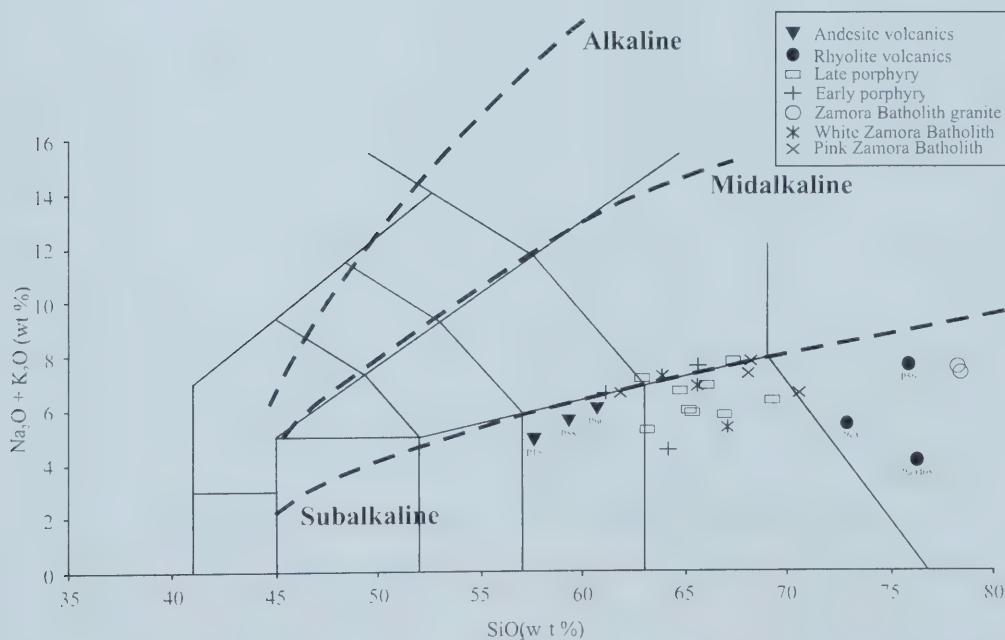


Fig. 6.1b Total alkali silica diagram with the fields for alkaline, midalkaline, and subalkaline rocks (after Rittmann, 1957) overlain on the classification fields of LeMaitre et al. (1989). Observe that the majority of the samples plot within the subalkaline field.

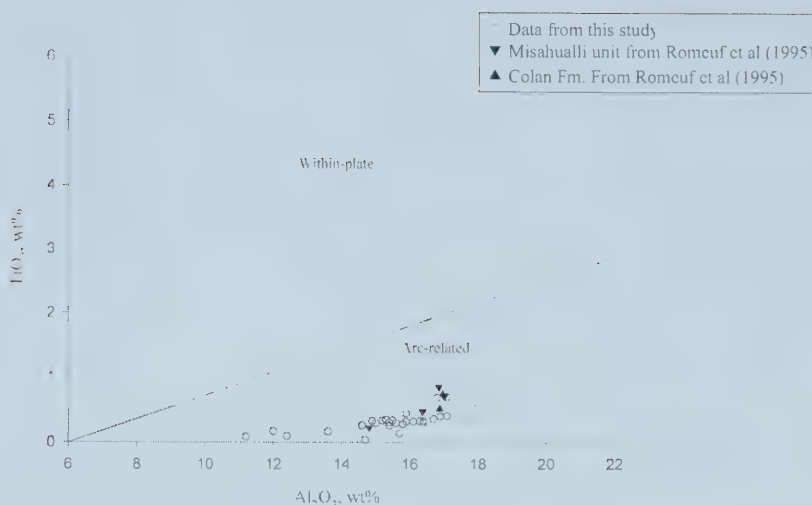


Fig. 6.2a. TiO_2 vs. Al_2O_3 diagram after Muller and Groves (1993) showing the discrimination fields between within-plate and arc-related magma series.

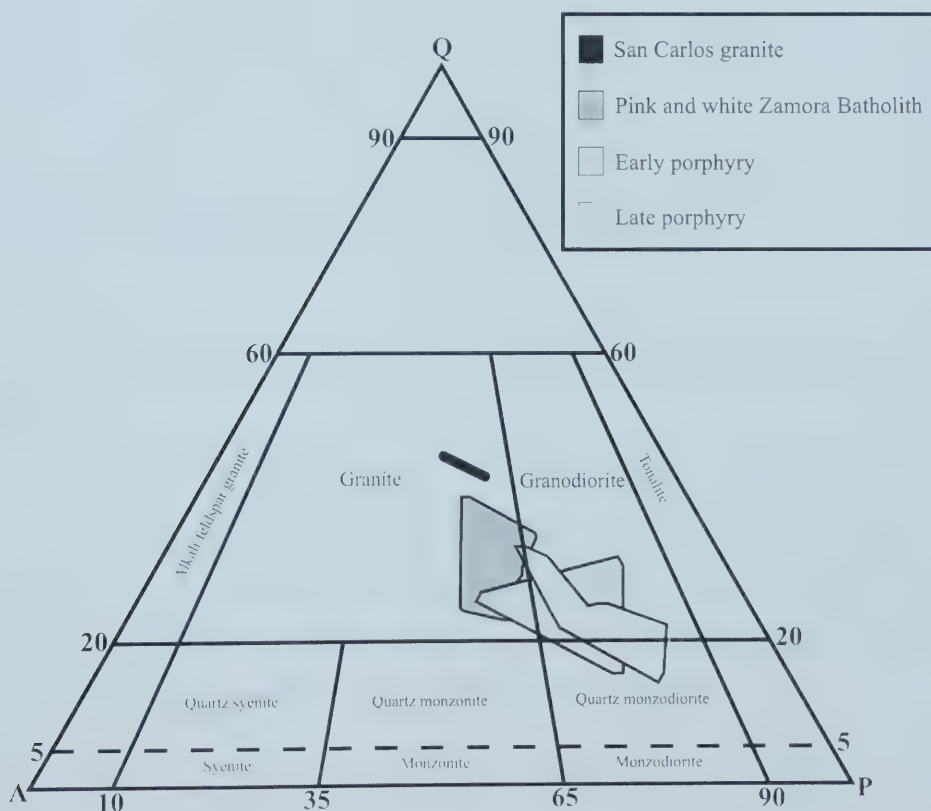


Fig. 6.2b. QAP ternary diagram after Streckeisen (1976) showing the classification of the intrusive rocks from the San Carlos and Panantza prospects.

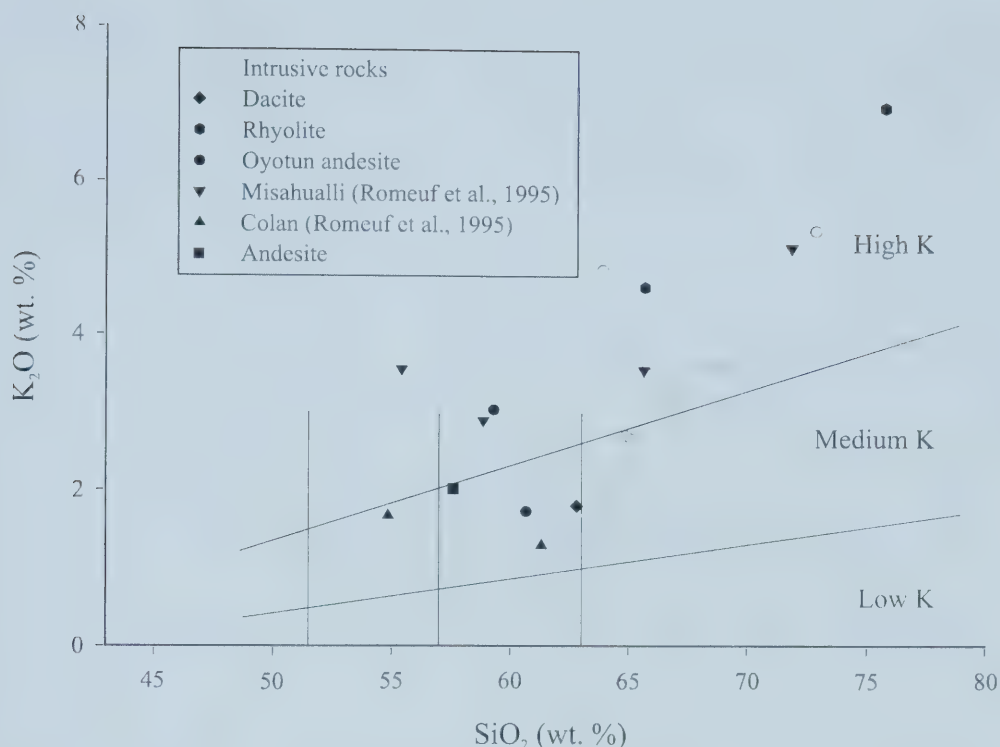


Fig. 6.3. K_2O vs. SiO_2 diagram after LeMaitre (1989) showing the medium to high K affinities for the sample suite with data from Romeuf et al. (1995).

Oxide-Oxide plots

Figure 6.4. shows major oxide vs. SiO_2 plots for the San Carlos and Panantza prospects.

Trace elements

Trace-element data are presented in Tables 6.1 and 6.2. Trace-element diagrams were constructed using the primitive mantle and CI chondrite normalizing values of Sun and McDonough (1989; Figs. 6.5a, 6.5b and 6.6). All of the samples present patterns indicative of a calc-alkaline volcanic arc and are discussed in Chapter 8.

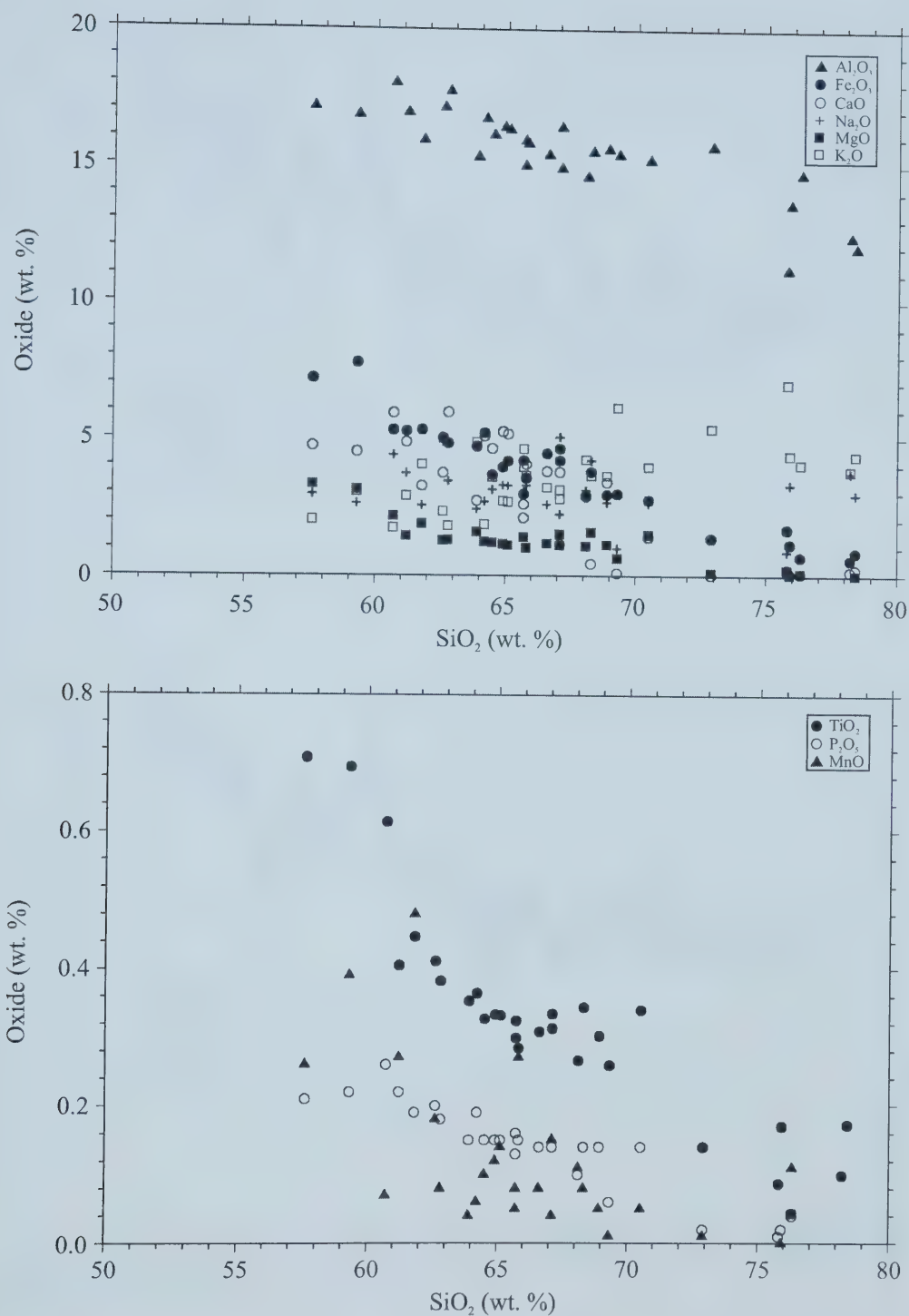


Fig. 6.4. Oxide versus SiO₂ plots for the San Carlos and Panatza prospects.

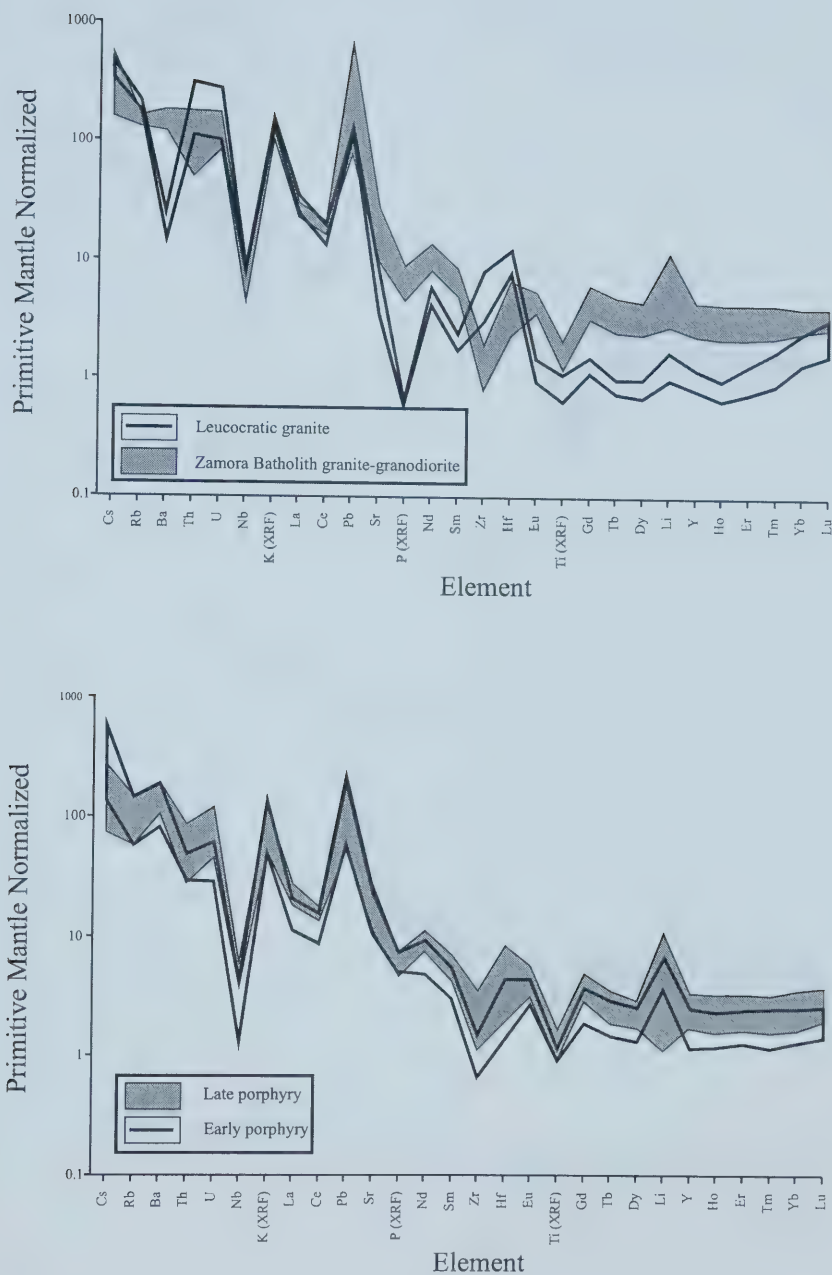


Fig. 6.5a. Primitive mantle normalized trace-element diagrams for the Zamora Batholith suite. Top: Leucocratic granite and Zamora Batholith showing the differences between the two units. Bottom: The early and late porphyry suites.

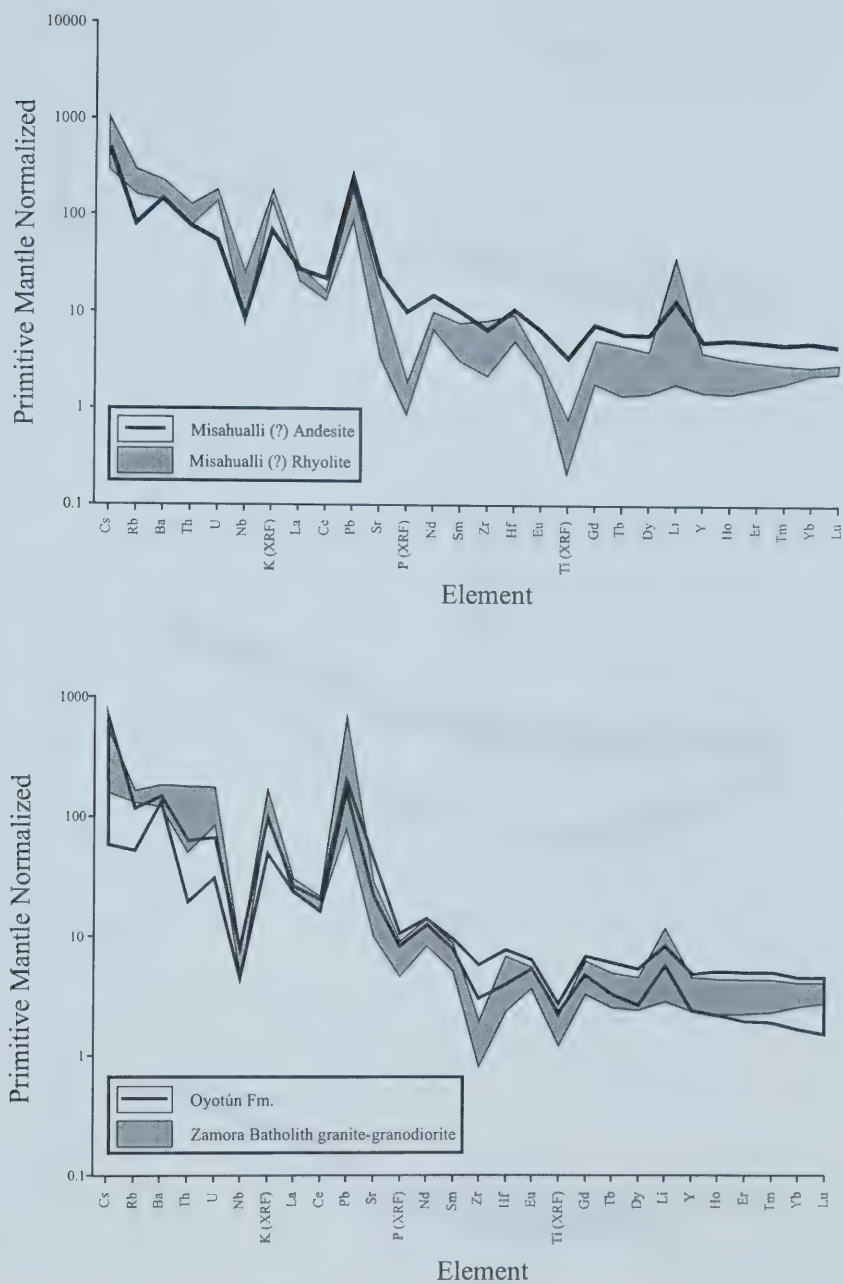


Fig. 6.5b. Primitive mantle normalized trace-element diagrams for the Zamora Batholith suite. Top: Misahualli (?) rhyolite and andesite. Bottom: The Zamora Btholith and Oyotún Fm. showing similar patterns.

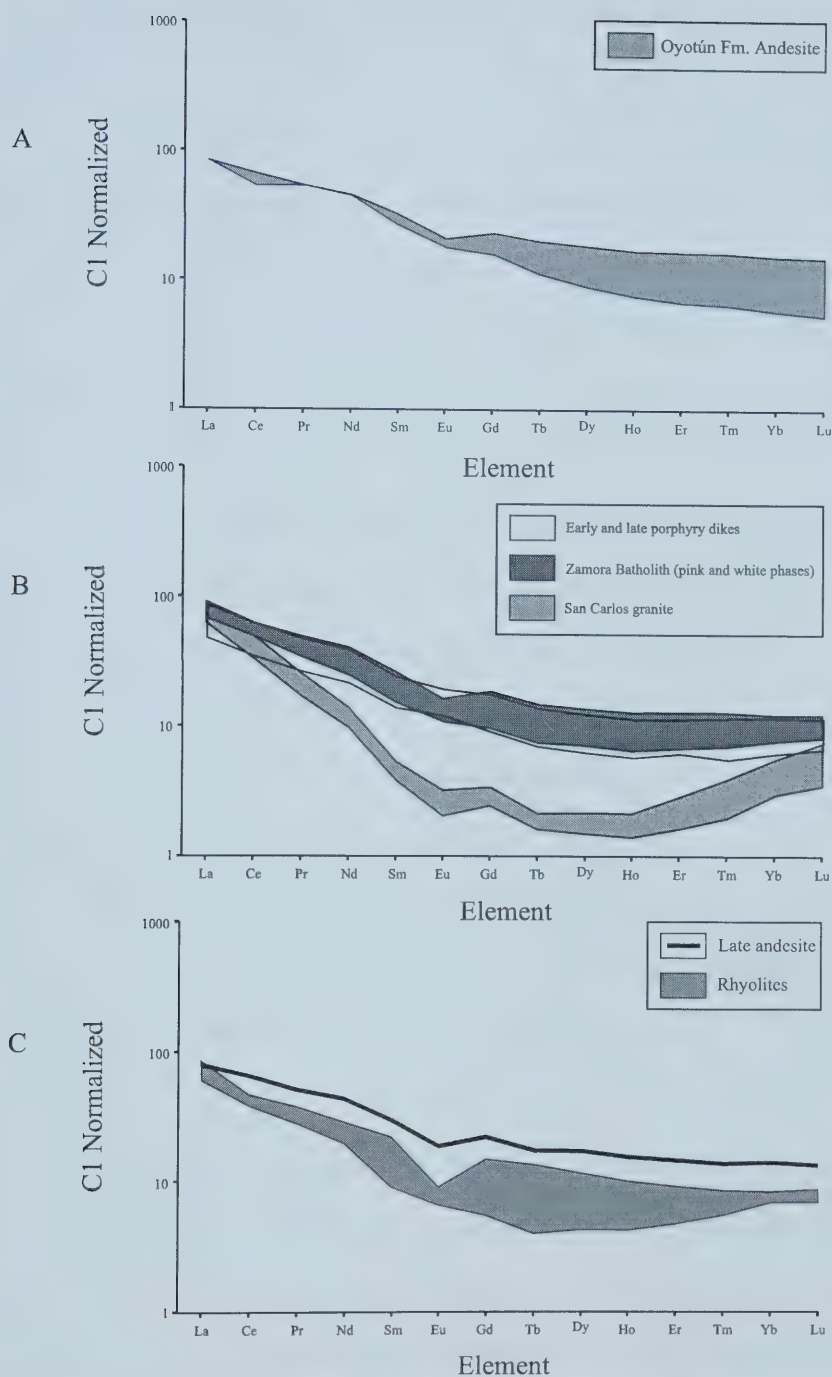


Fig. 6.6. Plots of rare earth element data of intrusive and volcanic rocks from the Zamora Batholith normalized to the C1 Chondrite composition of Sun and McDonough (1989). A. Oyotún Fm. andesite. B. Pink and white phases of the Zamora Batholith, early and late porphyry dikes, and San Carlos granite. C. Late andesite and rhyolite dikes.

7. Radiometric age-dating of the San Carlos, Panantza, and Sutsu prospects

Introduction

Thirty three samples from various phases of the Zamora Batholith have been dated previously using the K-Ar and Rb-Sr techniques, and provided ages ranging from 246 to 145 Ma (Baldock, 1982; Zamora and Litherland, 1993a). The combined results from K-Ar (Gendall et al., 2000), $^{40}\text{Ar}/^{39}\text{Ar}$ (this study), and Re-Os (this study) geochronology provide further constraints on the age of batholith emplacement, porphyry dike intrusion, and mineralization within the northern part of the Zamora Batholith.

The areas sampled and dated are the Kutucus, San Carlos, Panantza, and Sutsu prospects. All of the prospects with the exception of the Sutsu prospect are located within 5km of each other, whereas the Sutsu prospect is located 10km to the south (Fig. 4.3).

Sample Selection

$^{40}\text{Ar}/^{39}\text{Ar}$

Three surface samples from the San Carlos prospect were chosen for $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology. The samples were selected to provide an age for the Zamora Batholith, a late intermineral porphyry dike, and quartz-muscovite alteration.

K-Ar

Three samples were selected for dating by I. Gendall by the K-Ar method in order to obtain reconnaissance ages for the Kutucus and the San Carlos prospects. The Kutucus sample was taken from the contact between the Oyotún Formation and the Zamora Batholith. Both hornblende and biotite were dated. The San Carlos sample came from

drillhole 2 and was dated using the whole rock method; the results were previously reported by Gendall et al. (2000).

Re-Os

A single sample of molybdenite was selected from each of the three porphyry prospects to date the age of mineralization. Drillcore samples were used from the San Carlos and Panantza prospects whereas at the Sutsu prospect a surface sample was used.

Analytical Methods

Laser $^{40}\text{Ar}/^{39}\text{Ar}$ analyses were carried out at the Geological Survey of Canada laboratories in Ottawa, Ontario. Re-Os analyses were performed at the AIRIE high-level laboratory at Colorado State University, and the K-Ar analyses were performed at the Servicio Nacional de Geología y Minería (SERNAGEOMIN) in Santiago, Chile.

Principles of the Re-Os technique with respect to molybdenite

The sulfide mineral molybdenite (MoS_2) presents a unique case for the ^{187}Re - ^{187}Os (parent-daughter) radiometric dating method in that it is a sink for Re and essentially excludes Os from its structure upon formation. The Re concentrations for molybdenite are almost always well into the ppm range and may approach several percent substitution for Mo. In contrast, common or initial Os in molybdenite is generally in the very low ppt range. Thus, initial Re/Os ratios in molybdenite may easily exceed 10^6 . The Re-Os chronometer is based on the β^- decay of parent ^{187}Re (about 62% of total Re) to daughter ^{187}Os . Note that normal common Os contains approximately $\sim 1.5\%$ ^{187}Os , and because total common Os rarely exceeds the low ppt level, molybdenite effectively contains insignificant initial ^{187}Os for the purposes of age calculation. Because Re/Os ratios in molybdenite are exceedingly large, after only a very short period

of time measurable radiogenic ^{187}Os is present in the molybdenite, and a so-called model age can be determined from this single mineral sample. This age is obtained by applying the equation $^{187}\text{Os} = ^{187}\text{Re} (e^{\lambda t} - 1)$ where t is the age and λ is the decay constant for ^{187}Re . The $1.666 \times 10^{-11} \text{yr}^{-1}$ ^{187}Re decay constant of Smoliar et al. (1996) is used with an uncertainty of 0.31%.

Re-Os methodology

The Re-Os data in this study were obtained by the Carius tube method whereby the sample is dissolved and equilibrated with Re and Os spikes in $\text{HNO}_3\text{-HCl}$ (*aqua regia*)-sealed in a thick-walled glass ampule and heated for 12 hours at 230°C (Shirey and Walker, 1995). The Os is subsequently recovered by solvent extraction using CCl_4 .

Purified rhenium and osmium are loaded on Pt filaments and isotopic compositions are determined using negative thermal ion mass spectrometry (NTIMS; Creaser et al., 1991; Völkening et al., 1991). All chemistry was performed in the AIRIE high-level laboratory at Colorado State University. Samples were analyzed using the Faraday cup on a NBS 12-inch radius, 90° sector mass spectrometer. As needed, analyses were replicated to optimize sample/spike ratios.

Uncertainties in calculated ages primarily reflect the uncertainty in the ^{187}Re decay constant (0.31%) and the uncertainty in the ^{185}Re and ^{190}Os spike calibrations (0.05% and 0.15%, respectively). Uncertainties for single mineral ages on molybdenite range from ± 0.35 to $\pm 0.40\%$ at AIRIE. Re and Os blank corrections are insignificant. Recent blank determinations at the AIRIE molybdenite laboratory yielded total analytical Re blanks of 17-18 pg. and total Os blanks of 6-8 pg. with a variable $^{187}\text{Os}/^{188}\text{Os}$ ratio

ranging from 0.5 to 8.3. Also, two molybdenite standards of Archean and Permian age are routinely run at AIRIE, as an internal check (Stein et al., 1997; Markey et al., 1998).

⁴⁰Ar/³⁹Ar methodology

Three samples were selected for laser ⁴⁰Ar/³⁹Ar geochronology. Samples were prepared at the University of Alberta by crushing and sieving to recover the 120µm to 140µm fraction. Muscovite was crushed and separated using paper and static electricity. Further washing and ultrasonic cleaning removed any fine dust adhering to the sample. Heavy liquid separation of hornblende was accomplished using bromoform. The hornblende grains were then cleaned in acetone to remove any residual bromoform. Muscovite and hornblende grains were then hand picked under ethanol. The final separates of hornblende and muscovite grains were sent to the Geological Survey of Canada laboratories in Ottawa, Ontario, where they were loaded into aluminum foil packets along with a single grain of Fish Canyon Tuff Sanidine (FCT-SAN) to act as flux monitor (apparent age of 28.03 ± 0.1 Ma; Renne et al., 1994). These packets were placed, along with other samples bearing FCT-SAN, into six tubes arranged radially around an aluminum can 40 mm long $\times$ 19 mm diameter. Samples were irradiated for 8 hours at the research reactor of McMaster University in an approximate fast neutron fluence of 3×10^{16} neutrons/cm², with a resulting J-factor ranging from 0.00196400 to 0.00196980. Variation in J-factor is corrected for by utilizing the approximately 10 flux monitors per tube to interpolate along the length of the can.

Upon return from the reactor, samples were split into 2 aliquots, loaded into individual holes in a copper planchet, and heated and analyzed separately. Each aliquot was step heated in 3 to 11 increments, from 0.1 to 2 % of the nominal laser power, using

a Merchantek® MIR-10 10W CO₂ laser equipped with 2mm x 2mm flat-field lens, obviating the need for beam panning. Data were acquired with a VG3600 gas source mass spectrometer. Data collection protocols follow those detailed in Villeneuve and MacIntyre (1997), and data reduction procedures for individual aliquots follow Roddick (1988). Analytical results are presented in Table 1 and upon ascertaining reproducibility between aliquots the data were combined onto correlation plots (Roddick et al., 1980) for presentation and final data analysis. FCT-SAN has been demonstrated to be sufficiently homogeneous for laser ⁴⁰Ar-³⁹Ar dating (Baksi et al., 1996; Lanphere and Baadsgaard, 1997; Renne et al., 1994) and this has been confirmed by repeated analysis at the Geological Survey of Canada laboratory.

FCT-SAN is used as the flux monitor and is assigned an apparent age of 28.03 Ma (Renne et al., 1994). The combined uncertainty in flux gradient and natural variability between aliquots of FCT-SAN is estimated from the reproducibility of multiple analyses. All errors herein are presented at the 2σ level of uncertainty, unless otherwise noted. The data were combined onto side-by-side step heating gas release plots, normalized to the total volume of ³⁹Ar for both aliquots (Fig. 7.1). This plot retains the relative heating profiles produced, as well as giving a rough measure of sample size, for each individual aliquot. It also provides a visual means of comparing reproducibility of spectra between aliquots. Data reduction procedures for the combined plateau regions follow Roddick (1988) and are primarily dependent on the individual error associated with each step. In addition, the data are combined onto correlation plots (Roddick et al., 1980) for presentation and final data analysis. Error in the J-factor (± 0.5%, 1σ) is applied to the final age determination, but uncertainty in the age of the monitor is not.

Table 7. $^{40}\text{Ar}/^{39}\text{Ar}$ data

Power ^a	Volume ^{39}Ar $\times 10^{-11}$ cc	$^{39}\text{Ar}/^{39}\text{Ar}$	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{37}\text{Ar}/^{39}\text{Ar}$	$^{36}\text{Ar}/^{39}\text{Ar}$	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{40}\text{Ar}/^{39}\text{Ar}$	Apparent Age Ma
81 Muscovite; J=.00196980 (Z5590)										
Aliquot: A										
2.3**	0.47	0.0201±0.0016	0.009±722.718	0.011±0.001	47.406±0.856	12.6	41.452±0.772	4.9	141.60±2.54	
3.0	1.59	0.0055±0.0004	0.008±123.225	0.006±0.001	46.266±0.460	3.5	44.629±0.448	16.9	152.01±1.45	
3.4	1.52	0.0049±0.0007	0.006±231.887	0.006±0.001	45.946±0.410	3.2	44.490±0.408	6.1	181.56±1.37	
3.9	0.15	0.0322±0.0059	0.060±224.643	0.040±0.011	53.816±2.312	17.7	44.316±2.204	6	150.99±7.20	
12.0	0.62	0.0149±0.0012	0.011±205.189	0.014±0.002	48.843±0.704	9.0	44.428±0.650	6.6	151.36±2.12	
Aliquot: B										
2.3**	0.94	0.0117±0.0009	0.013±234.908	0.007±0.001	45.015±0.343	7.7	41.566±0.307	10.0	141.98±1.01	
3.0	1.93	0.0053±0.0004	0.005±313.850	0.002±0.001	46.137±0.528	3.4	44.574±0.524	20.5	151.83±1.71	
3.4	1.33	0.0059±0.0006	0.008±103.423	0.002±0.002	46.311±0.685	3.7	44.579±0.678	14.2	151.85±2.21	
12.0	0.86	0.0095±0.0009	0.011±135.391	0.005±0.001	48.005±0.822	5.9	45.189±0.799	9.2	153.84±2.61	
101E Hornblende; J=.00196650 (Z5591)										
Aliquot: A										
2.3**	0.05	3.2987±0.2496	1.787±113.831	2.072±0.222	1040.007±75.278	93.7	65.251±105.303	0.2	217.80±331.11	
3.0**	0.12	0.5981±0.0251	1.141±55.928	0.294±0.019	201.143±7.670	87.9	24.408±10.415	0.5	84.58±35.26	
3.4**	0.08	0.2733±0.0124	1.401±91.580	0.108±0.008	105.460±3.597	76.6	24.709±3.925	0.4	85.60±13.28	
3.9**	0.11	0.1904±0.0089	2.055±52.880	0.187±0.010	79.284±3.211	71.0	23.021±3.238	0.4	79.88±10.99	
4.6	3.69	0.0212±0.0002	8.6±5.121	1.102±0.005	53.576±0.184	11.7	47.311±0.183	15.2	160.49±0.59	
5.1	2.59	0.0181±0.0006	8.162±13.007	1.045±0.010	52.504±0.463	10.2	47.158±0.481	10.7	159.99±1.56	
5.8	2.92	0.0117±0.0006	7.423±12.941	1.207±0.012	50.754±0.427	6.8	47.302±0.447	12.0	160.46±1.45	
6.8**	0.64	0.0291±0.0019	6.949±35.181	1.083±0.035	49.189±1.155	17.5	40.575±1.209	2.6	138.50±3.97	
12.0**	0.89	0.0255±0.0012	6.708±21.230	0.842±0.012	44.775±0.613	16.5	37.232±0.630	3.7	127.48±2.08	
Aliquot: B										
2.3**	0.06	5.0377±0.1435	1.765±45.935	2.168±0.102	1588.615±39.581	93.7	99.976±57.817	0.2	323.81±171.42	
3.9**	0.25	0.5686±0.0174	1.825±44.810	0.337±0.018	202.978±4.926	82.8	34.966±7.029	1.0	119.97±23.33	
4.3**	0.34	0.1210±0.0030	6.823±11.776	0.662±0.022	77.822±0.718	45.9	42.075±0.796	1.4	143.42±2.61	
4.6	2.00	0.0249±0.0006	8.436±16.462	1.146±0.020	54.602±0.664	13.5	47.230±0.671	8.2	160.23±2.18	
5.1	2.96	0.0206±0.0008	9.138±10.275	1.096±0.008	53.649±0.356	11.4	47.556±0.417	12.2	161.29±1.35	
5.8	4.28	0.0157±0.0004	8.503±11.700	1.203±0.011	52.169±0.394	8.9	47.530±0.403	13.6	161.20±1.31	
6.8	1.53	0.0294±0.0018	7.659±23.118	1.154±0.022	55.667±0.831	15.6	46.972±0.969	6.5	159.39±3.15	
12.0**	1.76	0.0526±0.0008	7.352±14.903	0.987±0.014	59.387±0.415	26.7	43.829±0.451	7.3	149.16±1.47	
138 Hornblende; J=.0019640 (Z5592)										
Aliquot: A										
2.3**	0.06	2.3927±0.2975	2.187±182.724	1.012±0.189	738.560±89.312	93.7	5.530±125.268	0.2	108.40±417.97	
3.9**	0.30	0.2326±0.0055	2.495±31.162	0.202±0.005	103.839±1.490	66.2	35.112±1.998	1.0	120.31±6.62	
4.3**	0.66	0.0431±0.0015	6.387±36.149	0.651±0.017	55.242±1.275	23.1	42.505±1.281	2.2	144.65±4.19	
4.6	0.74	0.0666±0.0011	5.764±14.382	0.788±0.015	64.393±0.613	30.6	44.718±0.584	2.5	151.87±1.90	
5.1	1.55	0.0171±0.0014	6.368±16.410	0.818±0.015	49.987±0.537	10.1	44.938±0.654	5.1	152.59±2.13	
5.8	4.05	0.0119±0.0003	6.264±6.213	0.871±0.010	48.849±0.166	7.2	45.340±0.175	13.4	153.90±0.57	
6.8	2.74	0.0129±0.0003	6.107±7.193	0.875±0.007	49.030±0.241	7.8	45.225±0.236	9.0	153.52±0.77	
12.0**	1.96	0.0168±0.0006	6.305±14.625	0.673±0.009	48.728±0.447	10.2	43.762±0.464	6.5	148.76±1.51	
Aliquot: B										
2.3**	0.07	1.6185±0.0778	1.918±76.319	0.632±0.047	518.586±19.934	92.2	40.310±30.204	0.2	137.46±99.17	
3.9**	0.45	0.1841±0.0056	2.189±19.417	0.185±0.006	88.192±0.997	61.7	33.799±1.842	1.5	115.95±6.12	
4.5**	1.21	0.0314±0.0013	5.999±26.362	0.635±0.015	52.410±0.901	17.7	43.124±0.948	4.0	146.67±3.10	
4.6	2.13	0.0114±0.0004	6.215±14.922	0.770±0.011	47.905±0.458	7.0	44.541±0.458	7.0	151.29±1.49	
5	5.89	0.0080±0.0001	6.094±10.224	0.869±0.006	47.607±0.308	5.0	45.236±0.307	19.4	153.56±1.00	
5.8	5.73	0.0097±0.0001	6.118±7.252	0.847±0.005	47.949±0.226	6.0	45.084±0.225	18.9	153.06±0.73	
6.8**	0.97	0.0169±0.0008	4.080±32.056	0.533±0.013	44.593±0.890	11.2	39.586±0.877	3.2	135.08±2.88	
12.0**	1.84	0.0143±0.0007	4.489±14.281	0.567±0.011	44.605±0.277	9.5	40.367±0.304	6.1	137.64±1.00	

Notes to Table 7.1:

a: As measured by laser in % of full nominal power (10W)

b: Fraction ^{39}Ar as percent of total run

c: Errors are analytical only and do not reflect error in irradiation parameter J

d: Nominal J, referenced to FCT-SAN=28.03 Ma (Renne et al., 1994)

*: Radiogenic Ar

**: Not included in regression

ATM: Atmospheric Ar

All uncertainties quoted at the 2-sigma level

Geochronologic Results

⁴⁰Ar/³⁹Ar age results

Hornblende from the pink, medium-grained, equigranular granodiorite of the Zamora Batholith from the San Carlos prospect was selected for dating (sample 101E). The hornblende forms subhedral to euhedral crystals 2-5mm in length evenly dispersed within the rock. The grains were only very slightly chloritized and two aliquots were run. Aliquots 1 and 2 show a drop in apparent age in the highest temperature steps (6.8W and 12.0W; Table 7.1; Fig. 7.1a). This is probably due to degassing of a second phase, possibly minor chlorite. Aliquot 2 shows some Ar loss in the low temperature steps, which may indicate a partial thermal overprint; however, this was not reproduced with aliquot 1. Regressing the most radiogenic steps through the composition of atmospheric argon ($^{40}\text{Ar}/^{36}\text{Ar} = 295.5$) results in a date of 160.6 ± 1.6 Ma (MSWD = 0.6).

Slightly chloritized hornblende from a late granodiorite porphyry dike located near the center of the San Carlos prospect was selected for dating (sample 138). The hornblende forms euhedral crystals equally dispersed throughout the rock of between 5 and 7mm in length. Aliquots 1 and 2 both show Ar loss in the low temperature steps indicative of a slight thermal overprint. Both aliquots also show a drop in the apparent age in the highest temperature steps (6.8W and 12.0W; Fig. 7.1b) which, may indicate the degassing of a second phase, probably chlorite. Regressing the most radiogenic steps through the composition of atmospheric argon ($^{40}\text{Ar}/^{36}\text{Ar} = 295.5$) results in a date of 153.5 ± 1.5 Ma (MSWD = 1.8).

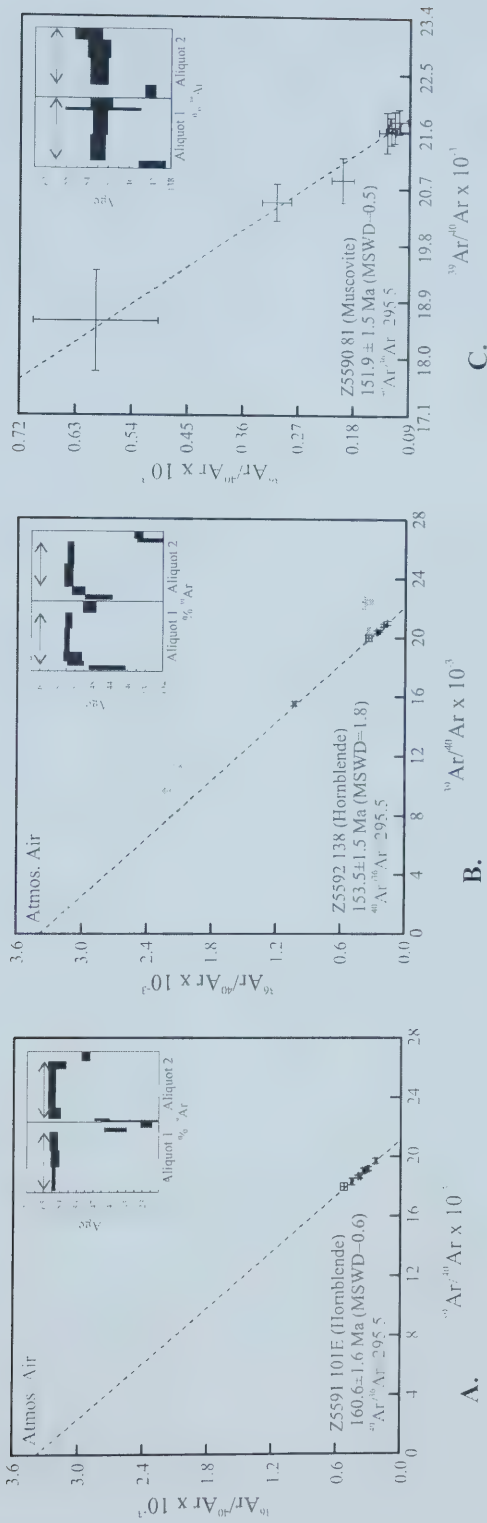


Fig. 7.1. Laser $^{39}\text{Ar}/^{39}\text{Ar}$ step-heating data for samples of hornblende and muscovite from the San Carlos prospect. Data are plotted on inverse-isochron diagrams from which regression ages have been calculated. Dark colored data points were used to calculate the regression, whereas light colored data points were not used. Gas-release spectra are also shown as insets and contain step heating data from two aliquots per sample, alternately shaded and normalized to 100% of the volume of ^{39}Ar released in each aliquot. The arrows located above the spectra indicate the steps forming a plateau. These spectra are included for visualization purposes because, although plateau ages can be calculated, regression of the inverse-isochron data provides a more accurate estimate of the age. A. Sample 101E, hornblende from the pink Zamora Batholith phase. B. Sample 138, hornblende from the late porphyry dike. C. Sample 81, muscovite from the matrix supported breccia.

Coarse-grained muscovite from a hydrothermal breccia located along the eastern edge of the San Carlos prospect was dated (sample 81; Fig. 7.1c). The muscovite forms subhedral to euhedral crystals between 0.5 and 1.5 mm in a quartz matrix with minor Cu and Mo sulfides. Aliquots 1 and 2 show a slight Ar loss in the first step, which may indicate a slight thermal overprint. Regressing the most radiogenic steps through the composition of atmospheric argon ($^{40}\text{Ar}/^{36}\text{Ar} = 295.5$) results in a date of 151.9 ± 1.5 Ma (MSWD = 0.5).

Re-Os age results

The results from the molybdenite samples are listed in Table 7.2. All of the samples contain somewhat elevated concentrations of Re which required adjustment of the spike ratios. The San Carlos prospect sample had to be run twice due to the discovery of elevated Re concentrations. The results of the two runs were similar, but the second date (157.0 ± 0.6 Ma) was preferred because of the spike problems in the first run. An apparent variability in the Re concentration is observed in this sample (743.45 and 831.66 ppm Re) for the two aliquots that were run. The lowest Re values came from the Panantza prospect (585.17 ppm) whereas the highest values came from the Sutsu prospect (1061.86 ppm).

$^{40}\text{K}/^{40}\text{Ar}$ age results

I. Gendall selected three samples for K-Ar dating in order to obtain a reconnaissance age on the San Carlos and Kutucus prospects. These data are presented in Table 7.3. These dates have been included in this study in order to compare the K-Ar system with the $^{40}\text{Ar}/^{39}\text{Ar}$ and Re-Os systems. The $^{40}\text{Ar}/^{39}\text{Ar}$ and Re-Os dates indicate that the K-Ar system was not reset.

Table 7.2. $^{187}\text{Re}/^{187}\text{Os}$ data.

Sample no.	AIRIE no.	Mineral	^{187}Re	^{187}Os	Date
			(ppm)	(ppb)	Ma $\pm$ 2s error
SC091198	CT-115	MoS ₂	831.66 (0.95)	1375.6 (2.7)	157.8 $\pm$ 0.6
SC091198	CT-135*	MoS ₂	743.45 (0.47)	1223.7 (2.0)	157.0 $\pm$ 0.6
PA020398	CT-136	MoS ₂	585.17 (0.41)	940.6 (1.5)	153.3 $\pm$ 0.5
Sut su	CT-137	MoS ₂	1061.86 (0.71)	1724.6 (2.6)	154.9 $\pm$ 0.5

* Duplicate analysis: preferred date.

Table 7.3. $^{40}\text{K}/^{40}\text{Ar}$ data.

Sample no.	Lab no.	Material	%K	^{40}Ar	%Ar	Date
					Atm	Ma $\pm$ 2s error
KZ-105	H390.KZ105	Hornblende	0.626	3.865	30	152 $\pm$ 5
KZ-105	H392.KZ105	Biotite	5.606	35.786	4	157 $\pm$ 4
SC02 148.60*	H544.SC	Whole Rock	3.908	24.49	2	154 $\pm$ 5

* Previously published in Gendallet al. (2000)

Atm: Atmospheric Ar

 ^{40}Ar : Total Ar

8. Discussion

Tectonic setting

Paleogeographic reconstructions of the Mesozoic volcanic arc between the Late Jurassic-Early Cretaceous show that it consisted of a partially emergent chain of islands emplaced into continental crust that progressively emerged from south to north, becoming dominantly subaerial in Colombia (Jaillard et al., 1990; Vila et al., 1996; Marschik and Fontbote, 1996).

The data from this study and from a study by Romeuf et al. (1995), who analyzed Jurassic volcanic sequences from Ecuador and northern Perú (the Misahuallí (?) and Colan Fms. respectively), are plotted on a Nb–Y discrimination diagram (Fig. 8.1a; Pearce et al., 1984). All of the data from both data sets fall within the volcanic-arc and syn-collisional granite field. Further separation of the tectonic setting is possible between these fields by using a Rb–(Y+Nb) diagram (Fig. 8.1b; Pearce et al., 1984). All data points including those from Romeuf et al. (1995), with the exception of a single sample from this study, fall within the volcanic-arc field and are consistent with previous observations of the volcanic arc being emplaced into continental crust.

Following the localized subaqueous eruption and deposition of the Late Triassic-earliest Jurassic Piuntza Member of the Santiago Fm. into a preexisting rift basin (Balkwill et al., 1995; Gendall et al., 2000), there was a period of volcanic quiescence that lasted throughout the Early Jurassic (~ 25 Ma.). This hiatus in volcanic activity suggests that a major change in the rate or angle of the subduction occurred (Kay et al., 1999; Tosdal and Richards, 20001). Similar gaps in the volcanic arc within the Andes are attributed to flattening of the subducting slab (James and Sacks, 1999; Kay et al., 1999).

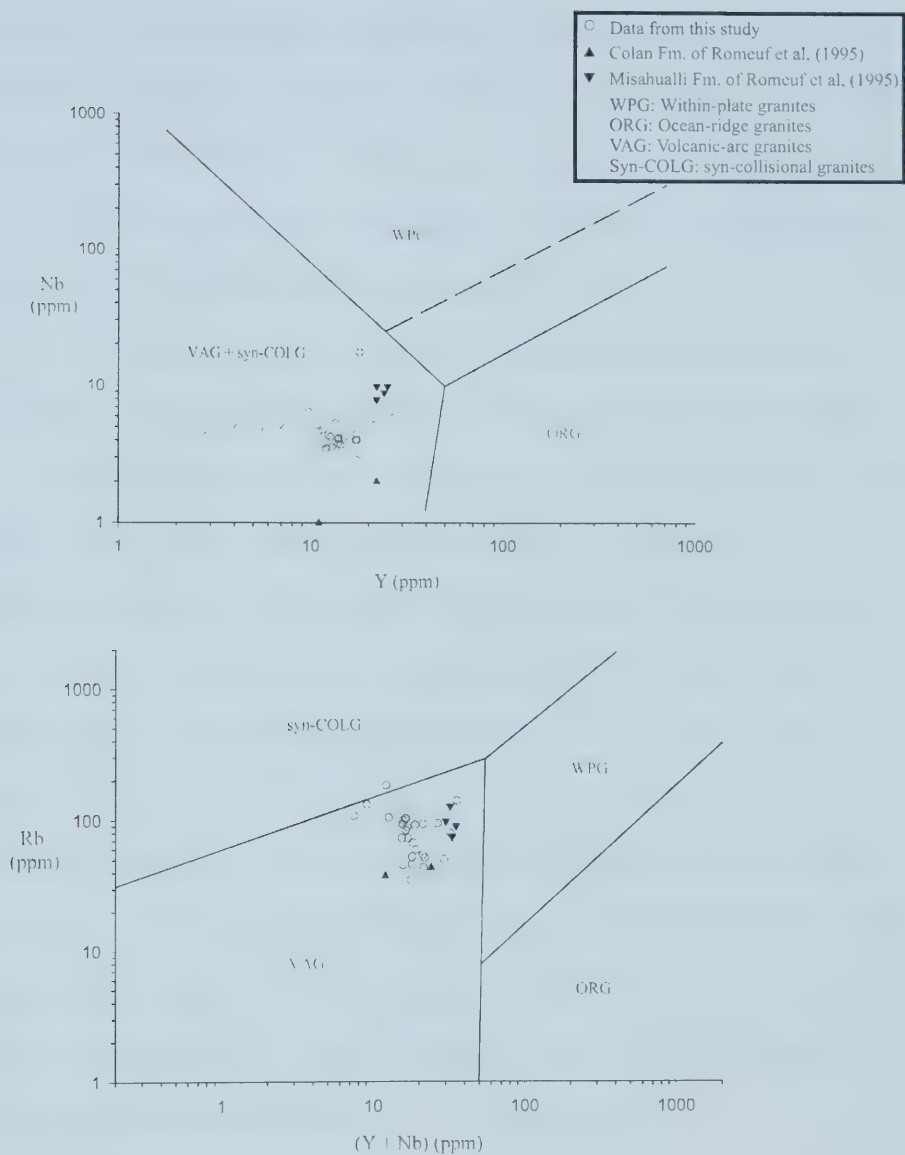


Fig. 8.1. A. Nb-Y tectonic discrimination diagram after Pearce et al. (1984) showing the data from this study with comparison data from Romeuf et al. (1995). B. Rb-(Y+Nb) diagram showing further separation of the tectonic setting with data from Romeuf et al. (1995).

Deposition of the carbonate Santiago Fm./Pucara Gp. continued, although deformation accompanied by variable uplift and erosion commenced as large amounts of magma began to intrude toward higher crustal levels (e.g., Zamora Batholith; Tschopp, 1953; Balkwill, 1995). This uplift created an erosional surface upon which the second volcano-sedimentary sequence was deposited (Quispesivana, 1996).

The widespread and prolonged subaqueous to subaerial volcanic activity (~ 15 Ma.) that began during the early Middle Jurassic and continued into the late Middle Jurassic, deposited the second volcanic sequence, the Oyotún Fm. (Tschopp, 1953; Quispesivana, 1996). The eruption of the Oyotún Fm. suggests that another reorganization of the angle of subduction, plate configuration, and stress regime occurred. Tosdal and Richards (2001) suggest that voluminous volcanic sequences erupt under a compressional stress regime whereas subvolcanic intrusion is favored during stress relaxation or tension.

Once the magmatic event that had erupted the Oyotún Fm. had terminated, magma began to pond at high crustal levels forming the Zamora Batholith. Cupolas of the Zamora Batholith intruded into the Oyotún Fm. and acted as the foci for the buildup of incompatible elements within the magma chambers. Periodic release of these fluids created the porphyry deposits, which are discussed in detail below.

Zamora Batholith

Deformation and uplift of the volcanic and sedimentary strata due to the growth of the Zamora Batholith occurred throughout the region. The steeply dipping faulted contacts observed for the Zamora Batholith (Zamora and Litherland, 1993a; INGEMMET, 2000) can be interpreted as preexisting extensional faults (Balkwill et al.,

1995; Gendall et al., 2000) that might have acted as feeder zones for magma, which suggests a large magma chamber at depth feeding into multiple subvolcanic magma chambers.

Porphyry copper mineralization in the Pangui District formed within cupolas of the Zamora Batholith that were intruded into the Oyotún Fm. An observation supported by the close spatial association of volcanic and intrusive rocks in the vicinity of many of the prospects (e.g., San Carlos, Panantza, Trinidad, San Miguel, and Warintza).

Intrusion of magmas to depths of <3 km is suggested by Tosdal and Richards (2001, citing Corry, 1988) to use a mechanism whereby the magma forces the overlying strata upward, causing deformation and faulting with lateral growth occurring by stoping or folding (Hamilton and Meyers, 1967; Paterson et al., 1996), the combination of which creates a laccolith.

Field evidence to support a subvolcanic level of intrusion for the Zamora Batholith in the study area is provided by Billiton geologists who observed vertically dipping sedimentary strata near the Kutucus prospect (Lassel, 1998; J. Ford pers. comm) suggesting that the Zamora Batholith in the study area deformed the overlying strata during its intrusion at 160 Ma. Furthermore, in the study area the Zamora Batholith stoped into the Oyotún Fm. (Fig. 5.3). The stoped contact and deformation suggest that the Zamora Batholith in the study area intruded to a subvolcanic depth of <3 km.

Geochemistry of the Jurassic igneous suite

Lithogeochemical results support the field evidence for the intrusion of the Zamora Batholith to subvolcanic depths. Primitive mantle normalized trace-element diagrams for all units show negative Nb and Ti anomalies and positive Pb anomalies

(Figs. 6.5a and 6.5b). C1 normalized rare earth element (REE) diagrams for samples of the Oyotún Fm. andesite show enrichments of light REE (LREE) but relatively flat middle- and heavy REE (MREE and HREE) patterns (Fig. 6.6a). Such trace-element and REE patterns are typical of calc-alkaline rocks (Rollinson, 1993) derived from relatively shallow sources in which garnet was not present (Frey et al., 1978; Hanson, 1980). If garnet had been present in the source, HREE would show a strong depletion relative to LREE and MREE. This is supported by an interpretation of a Bouguer anomaly map of Ecuador by Feininger and Seguin (1983) who modeled the continental crust underlying the Oriente region as 30-35 km thick.

The Zamora Batholith suite, with the exception of the leucocratic San Carlos granite, presents a very similar pattern to the Oyotún Fm. andesite: LREE enrichment but flat MREE and HREE patterns, again suggesting that garnet was not residual in the source region (Fig. 6.6b). The early and late porphyries present nearly identical patterns with the Zamora Batholith suite, suggesting that the porphyries may be genetically related to the Zamora Batholith (Fig. 6.6b). Moreover, the similar REE patterns for the Oyotún Fm. andesite, the Zamora Batholith suite, and the early and late porphyries suggest that all of the magmas were broadly cogenetic. The leucocratic San Carlos granite is distinctive, however, in showing strong depletion of MREE relative to LREE and HREE, suggesting that hornblende fractionation was important in the evolution of this magma (Fig. 6.6b; Hanson, 1978). It is therefore suggested that the leucocratic San Carlos granite differentiated from the Zamora Batholith, which contains abundant hornblende phenocrysts. The late andesite and rhyolite dikes show similar REE patterns to the Zamora Batholith suite (Fig. 6.6c).

Almost all of the data show weakly negative to positive Eu anomalies (Fig. 8.2a). Large negative Eu anomalies are often taken as an indication of plagioclase fractionation (Hanson, 1978), but in relatively oxidized calc-alkaline rocks, Eu is present as Eu^{3+} and is not extensively partitioned into feldspar (Hanson, 1978; Rollinson, 1993). The Oyotún Fm. andesite samples show a wide range with one sample having a barely positive anomaly and the other sample having a weak negative anomaly. The Zamora Batholith suite and associated porphyry rocks show weak to moderate negative anomalies with a single exception of an early porphyry sample, which is positive. The leucocratic granite also shows both weakly positive and negative anomalies. The late rhyolite and andesite show weak to moderate negative anomalies. Because all of the samples are feldspar phyrlic, these weak, variable anomalies are interpreted to reflect moderate magmatic oxidation states rather than the absence of plagioclase fractionation.

Plotting the data on a La_n/Sm_n vs. La_n/Yb_n diagram in order to elucidate possible fractionation trends shows that the data plot in a fairly restricted zone ($\text{La}_n/\text{Sm}_n = 2-5$ and $\text{La}_n/\text{Yb}_n = 5-14$ respectively; Fig. 8.2b). There are two exceptions, the granite ($\text{La}_n/\text{Sm}_n = 17$ and $\text{La}_n/\text{Yb}_n = 15-20$) and the rhyolite ($\text{La}_n/\text{Sm}_n = 2-9.5$ and $\text{La}_n/\text{Yb}_n = 7-12$).

Interpretation of Figure 8.2b suggests that the low ratios of La_n/Yb_n indicate that source of the Zamora Batholith suite was under relatively low pressures (i.e., crustal depths of <30-35 km) in the absence of residual garnet but that the leucocratic granite was under somewhat higher-pressure (i.e., a greater crustal depth of >35-55 km; Kay, 1991). However, Kay et al. (1999) noted that if the source of the rocks was at high-pressures, a decrease in the negative Eu^* anomaly along with increasing Na_2O and Sr content should be observed due to plagioclase breakdown. However, the leucocratic

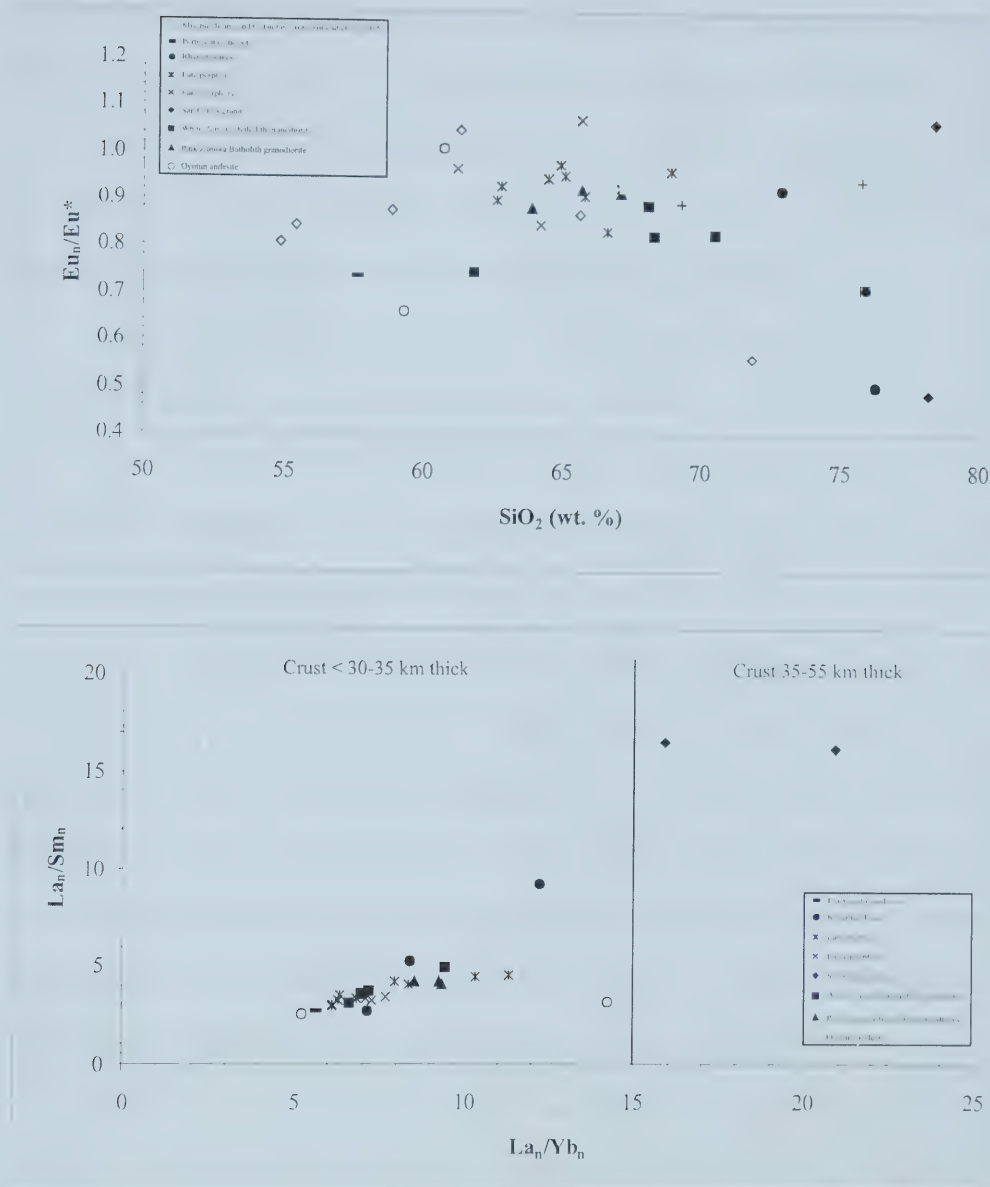


Fig. 8.2. A. Plot of Eu-anomaly versus SiO_2 for the Zamora Batholith suite of rocks with additional data from Romeuf et al. (1995). ($Eu_n/Eu^* = Eu_n / (Sm_n \times Gd_n)$; CI normalization values of Sun and McDonough, 1989) B. Plot of La/Sr versus La/Yb ratios of the Zamora Batholith rocks. Boundary at $La/Yb = 15$ is suggested by Kay (1991) to correspond to approximately 35 Km crustal thickness.

granite samples show the lowest values for Na_2O (≤ 0.04 wt. %) and Sr (≤ 140 ppm) versus averages of 1.0 wt. % for Na_2O and average of 500 ppm Sr for the rest of the sample suite. These observations are inconsistent with a model in which the San Carlos leucocratic granite fractionated at greater depth than the Zamora Batholith suite and may be simply explained by more extensive fractionation at low pressures.

The data were also plotted on a Th vs. $1/\text{Yb}$ diagram to ascertain the extent of crustal contamination (increased Th) versus garnet fractionation ($1/\text{Yb}$; Fig. 8.3; Richards and Villeneuve, in prep.). All of the samples with the exception of the leucocratic granite contain $1/\text{Yb}$ values of $<1 \text{ ppm}^{-1}$ whereas the leucocratic granite plots at 1.05 and 1.96, again suggesting that garnet fractionation was not important. In contrast, relatively high Th values (>10 ppm in some samples) suggest that crustal contamination of magmas was important (Feeley and Davidson, 1994). The Oyoútún Fm. andesite, early and late porphyry, and late andesite samples have lowest Th values (2-7 ppm). The Zamora Batholith suite shows some variability (4-16 ppm) and no separation between the pink and white phases can be made. The late rhyolite samples plot with the other samples, although they have a slightly higher range of Th on average (8-11 ppm). The leucocratic granite shows wide variability (9-26 ppm) and plots outside of the group.

The low Th values suggest that the more mafic samples (i.e., Oyoútún Fm. andesite, early and late porphyry, and late andesite samples) show the least evidence of crustal contamination, whereas the Zamora Batholith suite and more felsic rocks show an increase in Th content suggesting greater amounts of crustal contamination.

The relative lack of crustal contamination within the Oyoútún Fm. andesite suggests that the magma had a rapid ascent to the surface and spent relatively little time

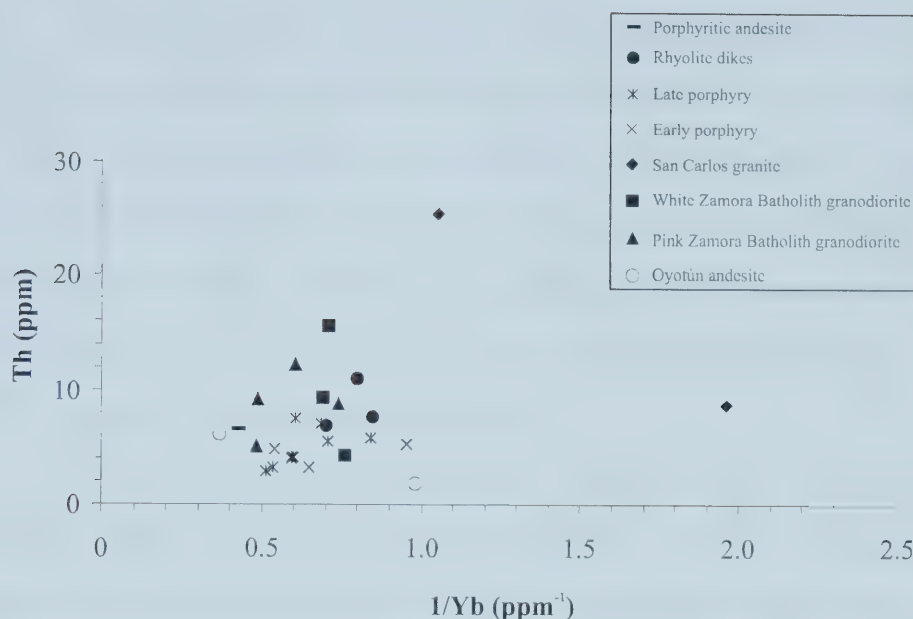


Fig. 8.3. Plot of Th versus $1/Yb$. (After Richards and Villeneuve, in prep). > 10 ppm Th is suggested by Feeley and Davidson (1994) as showing important amounts of crustal contamination.

in subvolcanic magma chambers. This is consistent with the suggestion by Tosdal and Richards (2001) that magmas in a compressional stress environment will only erupt if under high-pressure and suggests a compressional stress regime for the region during the Early-Middle Jurassic. However, the somewhat younger Zamora Batholith granite-granodiorite does show some evidence of crustal contamination which is consistent with a magma ponding in a high level magma chamber and assimilating country rocks (i.e., Paleozoic meta-sediments, Oyotún Fm. andesite and Jurassic clastic sediments). The slightly more mafic composition of the early and late porphyries when compared to the Zamora Batholith suggests that periodic magma recharge may have occurred.

The geologic history, field observations, and lithogeochemistry indicate that the magmas that comprise the Zamora Batholith were hydrous (e.g., hornblende and biotite bearing, suggesting a relatively shallow source), were intruded into relatively thin continental crust (< 35 km thickness), fractionated at shallow (subvolcanic) levels, and that the magma chambers underwent periodic recharge.

All of these factors are necessary in the formation of porphyry copper deposits, but without a favorable stress regime facilitating upper crustal magma emplacement these deposits would not have formed (Tosdal and Richards, 2001). The sequences of eruption and intrusion of the igneous rocks suggest repeated changing of the stress regime, which during the Late Jurassic appears to have provided a favorable environment for the subvolcanic intrusion of magma and hence porphyry copper formation (see below).

Jurassic tectonics and porphyry copper formation

The intrusion of the Zamora Batholith to a subvolcanic level in the study area coincided with a hiatus in volcanic activity that lasted (~ 25 Ma) into the Late Jurassic and suggests a reorganization of the stress regime. Right lateral movement along faults in Ecuador and northward movement of the Chaucha and Amotape-Tahuin terranes prior to their accretion to the margin of NW South America during the Early Cretaceous has been documented by various authors (e.g., Mourier et al., 1988; Zamora and Litherland, 1993a; Fig. 2.4) and suggests a transform fault may have existed during the Late Jurassic. If this were the case, a transtensional or transpressional stress regime would likely have dominated the region.

Tosdal and Richards (2001) suggest that such oblique stress regimes provide the maximum potential for porphyry copper deposit development as space can be made for subvolcanic magma emplacement at extensional pull-aparts and fault jogs. Indeed it is possible that pressure build up within the magma chambers was periodically released without major eruptions, an observation supported by the porphyritic textures of igneous rocks observed in the study area.

Porphyry copper formation in Colombia and Ecuador also occurred at this time (Sillitoe et al., 1982; Sillitoe, 1984; Gendall et al., 2000). The porphyry copper deposits within the Colombia-Ecuador segment of the volcanic arc appear to have formed between ~170 and 129 Ma (Sillitoe et al., 1982; Sillitoe, 1988; Gendall et al., 2000; this study). K-Ar dates from porphyry prospects in Colombia show alteration events at 166 ± 4 , $146-144 \pm 3$, and 131 ± 2 Ma (Sillitoe et al., 1982). Although the data are somewhat ambiguous, they do indicate three periods of porphyry emplacement within Colombia. Geochronologic data from Ecuador show porphyry emplacement occurred between ~160-150 Ma and two or three porphyries associated with mineralization are observed at the prospects.

Continued uplift caused erosion of material from the volcanic edifice to be shed eastward into the back arc basin depositing the clastic red-bed Chapiza Fm. over the Santiago and Oyotún Fms. (Tschopp, 1953; Zamora and Litherland, 1993a; Balkwill et al., 1995; Quispesivana, 1996) indicating widespread subaerial exposure of the volcanic edifice and later, the back arc basin.

Volcanic activity did not recommence until the Latest Jurassic–Early Cretaceous when the Misahualli Member of the Chapiza Fm. was deposited. The eruption of basalts was followed by rhyolite tuffs and coincides with the docking of the Chaucha and Tahuin–Amotape terranes (Hall and Calle, 1983; Zamora and Litherland, 1993a; Noble et al., 1997). The docking of the terranes caused the trench to jump westward. This appears to be consistent with observations by Noble et al. (1997) who noted a disturbance of the K-Ar and Rb-Sr systems in the region that correspond to a period of deformation and metamorphism between 120-140 Ma.

The Misahualli Member represents the last volcanic activity of the Jurassic volcanic arc which was active from the Latest Jurassic into the Early Cretaceous; after this time the region rapidly subsided. Uplift to the east began as the south Atlantic began to open during the Aptian–Albian (119-97 Ma) and the Hollin Fm. was deposited unconformably over the older strata.

The poor definition of the three volcanic sequences has led to some confusion over the names of the sequences, particularly in Ecuador. The most common usage is to use the Misahualli as a formation that combines all three volcanic sequences (Zamora and Litherland, 1993a; Balkwill et al., 1995; Romeuf et al., 1995), although by definition it is a member of the Chapiza Fm. (Tschopp, 1953). In Ecuador the older two of the three volcanic sequences are members of the Santiago Fm./Pucara Gp. and therefore cannot be the Misahualli Member of the Chapiza Fm. On the geologic map of Ecuador, Zamora and Litherland (1993a) combined the Chapiza Fm. and its Misahualli Member into the Santiago Fm., extending the age of the Misahualli Member (?) from the Early to Middle Jurassic (~ 185-150 Ma). However, Hall and Calle (1982) provide evidence that the Misahualli Member is Early Cretaceous, citing paleontological data and a K-Ar age of 132 Ma. Although the K-Ar age could be reset, it appears likely that the date is correct as it is supported by paleontological data. Furthermore the literature review shows that three distinct volcanic sequences exist.

Clarification of the relationship between stratigraphic units is important from an exploration perspective. Because the three distinct volcanic sequences were erupted under different conditions and into different environments, each volcanic sequence can be expected to be prospective for different types of mineral deposits.

The Piuntza Member of the Santiago Fm. hosts no known mineral deposits temporally associated with it (Zamora and Litherland, 1993b). Only later, during the intrusion of the Zamora Batholith did the Piuntza Member become mineralized with Late Jurassic (?) subeconomic base metal skarn deposits (Gemuts et al., 1992; Rodgers, 1998; Gendall et al., 2000). During the Late Cretaceous-Early Tertiary, low sulfidation Au mineralization overprinted the older base metal mineralization (Gemuts et al., 1992; Rodgers, 1998; Gendall et al., 2000).

The Oyoútún Fm. is spatially associated with substantial porphyry copper mineralization hosted within cupolas of the Zamora Batholith and hosts subeconomic base metal skarn deposits (e.g., Kutucus, Nambija). Equivalent stratigraphic sequences in Colombia and Chile also host porphyry and manto mineralization (Sillitoe et al., 1982; Espinoza et al., 1996; Vila et al., 1996; Marschik and Fontboté, 1996; Sillitoe, 1988). Potential for other types of mineralization exists within the Oyoútún Fm. from younger magmatic events (e.g., the Nambija District).

The Misahualli Member of the Chapiza Fm. is not known to host mineralization (Zamora and Litherland, 1993b). However, potential exists for Au deposits of latest Jurassic-Early Cretaceous age associated with the Misahualli Member.

Metallogeny

Figure 4.1 shows volcanic arcs of Jurassic through Neogene ages cross in the vicinity of the town of Huancabamba. North of Huancabamba the arcs young to the west, whereas south of Huancabamba the arcs young to the east. This apparent reversal of the age of the arcs can be explained by differences in the geologic history between the northern third and southern two thirds of South America. Multiple terrane accretions

along the margin of the northern segment caused repeated cessation of subduction and movement of the trench to the west (Toussaint and Restrepo, 1982; Aspden et al., 1987; Feininger, 1987; Mourier et al., 1988; Zamora and Litherland, 1993a; Arculus et al., 1999). The multiple terrane accretions may also account for the gap in the Cretaceous arc, which suggests that the angle of the subducted slab was flat and that a steep angle of subduction was not reestablished until some time during the Paleogene (James and Sacks, 1999; Kay et al., 1999).

In contrast the southern segment did not undergo terrane accretion but instead underwent tectonic erosion which has preserved the linear belts which migrated eastward as the margin was tectonically eroded and the subducting slab flattened out (Sillitoe, 1988).

Mineralization within the northern and southern segments mirrors this spatial and temporal relationship (Sillitoe, 1988; Zamora and Litherland, 1993b). The recycling of older material has been suggested by Sillitoe (1988) and may have implications for the formation of the world class porphyry deposits in the southern segment. Because the Jurassic arc has widespread and some of the oldest well documented Cu mineralization it is conceivable that large amounts of Cu were subducted and recycled into Tertiary deposits (Sillitoe, 1988). If this were the case, the northern segment would not have undergone the same recycling of Cu, suggesting that large size deposits containing recycled material would not be present in the younger arcs. This is supported by the lack of known large porphyry deposits within the accreted terranes (Zamora and Litherland, 1993b). Although the northern segment appears to lack potential for recycling of copper, the fact that multiple volcanic belts cross in the vicinity of the town of Huancabamba

suggests that potential exists for multiple ages of mineralization to have been overprinted. An example of this is observed in the Cordillera del Condor, where the Cretaceous-Tertiary Nambija low sulfidation Au District partially overprints the Jurassic Pangui Cu porphyry District (Gendall et al., 2000).

Pangui District Porphyry Model

The Pangui District porphyries were emplaced near the contact between the Zamora Batholith and the overlying Oyoútún Fm. They contain high hypogene copper grades and are large multi-pulse hydrothermal systems, which resulted in overprinting of the alteration zones, both vertically and laterally. Radiometric age dating and observations of field relationships from the San Carlos prospect and observations of field relationships at the Panantza and Wawame/Mirador prospects show a pattern where each younger pulse of mineralization is slightly offset to the SE and somewhat deeper than the preceding pulse (Blair, 2000; Gendall et al., 2000). This suggests fault control was important for the localization of the mineralization and is consistent with the mapped secondary faults that strike NW-SE in the region (Zamora and Litherland, 1993a).

The type of alteration observed in outcrop can be used to assess the erosion level of the porphyry deposit. At high levels of exposure (low erosion levels), porphyry deposits within the Pangui District show phyllic alteration with minor remnant potassic alteration and a variable amount of a structurally controlled intermediate argillic overprint (Blair, 2000; Gendall et al., 2000). Intermediate levels of exposure are dominated by potassic alteration with variable amounts of phyllic and intermediate argillic overprinting. The deepest levels of exposure will contain well-developed potassic alteration with disseminations and veinlets of magnetite, particularly in the more gold

rich systems (e.g., Panantza). If advanced argillic alteration formed during the emplacement of the porphyries it has now been eroded. Laterally, the deposits show a considerable amount of variation in the alteration due to the overprinting (e.g., San Carlos and Panantza).

Brecciation occurs at most of the prospects but so far it has not been known as a primary focus for ore deposition. Sillitoe and Sawkins (1971) suggest that the existence of breccia pipes indicate close proximity to the paleosurface, which is consistent with previous observations. In the Pangui District the breccia pipes also host mineralization, both as mineralized clasts and as open space fillings. The best documented example of this is the Wawame/Mirador prospect where an intramineral breccia with well-developed quartz-sericite-pyrite alteration hosts economic copper grades and potassic alteration bearing purple anhydrite at depth with molybdenite (Blair, 2000). Panantza also hosts similar purple anhydrite $\pm$ molybdenite veins and San Carlos hosts an untested quartz-sericite-pyrite $\pm$ chalcopyrite breccia very similar to the one described from Warintza/Wawame.

The last pulses of alteration/mineralization at the San Carlos, Panantza, and possibly Wawame/Mirador prospects appear to be related to breccia pipes and host phyllic alteration. This may indicate that the time between the first two pulses of mineralization was not long enough to allow the hydrothermal fluids to cool and become more acidic (Hedenquist and Richards, 1998; Tosdal and Richards, 2001). However the third or last pulse did create a phyllic alteration zone which suggests that the hydrothermal fluid did have time to cool.

Generally the Pangui District porphyries are pyrite-poor and dominated by chalcopyrite with subordinate molybdenite $\pm$ gold in some of the porphyries (e.g., Panantza and Wawame/Mirador; Blair, 2000; Gendall et al., 2000). Most sulfides appear as disseminations replacing primary mafic minerals, and although veins and veinlets do exist they are subordinate.

The overprinting of the mineralization/alteration has caused local remobilization of both copper and gold. This is best exemplified by the Panantza, San Carlos, and Wawame/Mirador prospects where the slight overlap of the alteration/mineralization events has caused local depletion or enrichment of gold and copper (Blair, 2000; Gendall et al., 2000; this study). The enrichment occurs in the younger alteration, whereas the depletion affects the older alteration/mineralization.

Late Dacite porphyry

The deposits in Colombia and Ecuador commonly host late dacitic porphyry intrusive events. The dacites vary in composition from rhyodacite to dacite and they appear to have been emplaced during the Early Cretaceous (Sillitoe et al., 1982, 1984; Blair, 2000; Gendall et al., 2000). These porphyries in Colombia were dated by Sillitoe et al. (1982, 1984) who dated an intramineral, sericite-altered, dacite porphyry at Infierno-Chili at 131 ± 2 Ma, and a magnetite-rich, slightly altered, late mineral, dacite porphyry from Mocoa at 136 ± 3 Ma. Field observations show that an early Cretaceous dacite-rhyodacite porphyry intrusive event also occurred within the Pangui District of Ecuador. There, post mineral dacite-rhyodacite porphyries intrude the deposits and the lower strata of the Hollin Fm. Although they it has not been dated by radiometric methods, the Hollin Fm. is known to be Early Cretaceous in age providing a youngest age of ~ 110 Ma for the

porphyries. Locally these porphyry dikes are altered either to a clay-carbonate or a pyrite-magnetite assemblage, similar to the observations of Sillitoe et al. (1982; 1984) from Colombia.

Because porphyry copper mineralization occurs with dacite porphyries at the Infierno-Chili prospect in Colombia, it appears likely that this type of mineralization could have formed within the Pangui District. Fieldwork has shown that alteration (silicification) and minor mineralization (pyrite) exist within the Hollin Fm. A float sample of a matrix-supported breccia whose clasts consist of angular pieces of the Hollin Fm in a matrix of Fe-Mn oxides was found in the Rio Panantza and may indicate the existence of breccia pipe mineralization. Within the Pangui District, native gold has been panned from the streams and rivers, the source of which has been interpreted by I. Gendall to be the unconformity between the Hollin Fm and the Zamora Batholith. However, an alternative interpretation for the source of the gold could be the dacite porphyries or a deposit type exemplified by the ones in the Angasmarca gold district, located south of Cajamarca in Peru (Noble and McKee, 1999). The Angasmarca district is hosted in the Goyarisquizca Group (correlative to the Hollin Fm in Ecuador) and consists of a structurally controlled group of hydrothermal deposits with arsenopyrite-pyrite and native gold emplaced during the Miocene (Noble and McKee, 1999). If the gold mineralization is from an Angasmarca type deposit, then it may indicate that Miocene mineralization overprints both the Pangui copper and Nambija gold districts.

Supergene enrichment

Both prospects show well-developed Cu oxide mineralization and the local development of a subordinate chalcocite enrichment blanket. Chávez (2000) suggests that

porphyry systems with low total sulfide volumes will produce only modest quantities of acid-sulfate bearing solutions, that can leach primary Cu sulfides. Copper may then be reprecipitated from these fluids as secondary sulfides and oxides, when and where the fluids are neutralized by mixing with groundwaters or reacting with feldspars. The lack of abundant pyrite associated with the ore shells of the porphyries dictates in-situ oxide development because the potassic alteration zones effectively neutralize the acidic fluids before they can move any great distance (Chávez, 2000; Gendall, 2000). The grade and tenor of the oxide zone is generally consistent with the hypogene grades although some enrichment has occurred. There is an upper leached zone dominated by Fe-Mn oxides and clay. Below this Fe-Mn oxide zone is a zone of clay and copper oxides that give the rock a green-blue color which in turn is followed by a zone containing locally developed secondary sulfide minerals (chalcocite and covellite).

The oxides are better developed at the Panantza prospect where they form a 500 x 300m blanket of variable thickness centered on the overlap between the two potassic alteration centers (Fig. 5.10). The oxides appear to have been sourced from the early potassic center, which contains a higher pyrite/magnetite to chalcopyrite ratio and is situated upslope from the younger potassic center. With the abundant rainfall in the region, oxidation of the sulfides occurs quite readily and downslope transport of acidic copper enriched water occurs. Upon reaching the second potassic center the solution is neutralized and the Cu drops out as oxides.

9. Conclusions

The porphyry copper deposits within the Colombia-Ecuador segment of the volcanic arc appear to have formed between 170 and 150 Ma (Sillitoe et al., 1982; Sillitoe, 1988; Gendall et al., 2000; this study).

Lithogeochemical data show that the Zamora Batholith, porphyry phases, and Oyotún Fm. are broadly cogenetic. It also appears that the Zamora Batholith intruded its own volcanic pile and it is suggested that the Chapiza Fm. was deposited from sediments shed off of the volcanic pile during a period of rapid uplift and erosion that corresponded with porphyry copper mineralization. Later volcanic activity in the region deposited the Misahualli Member of the Chapiza Fm. This indicates that the Pangui District is not hosted within the Misahualli Member but is hosted within the Oyotún Fm.

Potential exists within the region for at least two and possibly three ages of mineralization and two or more deposit types to have overprinted each other (e.g., Pangui porphyry copper – Nambija low sulfidation Au – breccia Fe-Au?).

Two pulses of mineralization and alteration are evident at the Panantza prospect whereas three pulses are observed at the San Carlos prospect. Field relationships and radiometric age determinations suggest that the pulses came in approximately 2-3 m.y. apart. The earliest pulse is located to the NW and the second/third pulse/s while slightly overlapping the first, are located to the SE.

The new Re-Os (157.0 ± 0.6 and 153.3 ± 0.5 Ma) and $^{40}\text{Ar}/^{39}\text{Ar}$ (160.6 ± 1.6 , 153.5 ± 1.5 , and 151.9 ± 1.5 Ma) dates confirm an imprecise $^{40}\text{K}/^{40}\text{Ar}$ date of 154 ± 5 Ma obtained from drillhole 2 at the San Carlos prospect by Gendall et al. (2000).

Supergene enrichment at San Carlos and Panantza can be expected to consist of in-situ oxides with local development of secondary sulfide minerals. The oxide development is dictated by the relative lack of pyrite and the widespread potassic alteration zones, which neutralize the acidic fluids developed during oxidation of sulfide minerals.

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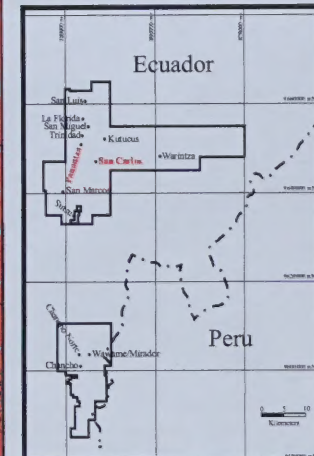
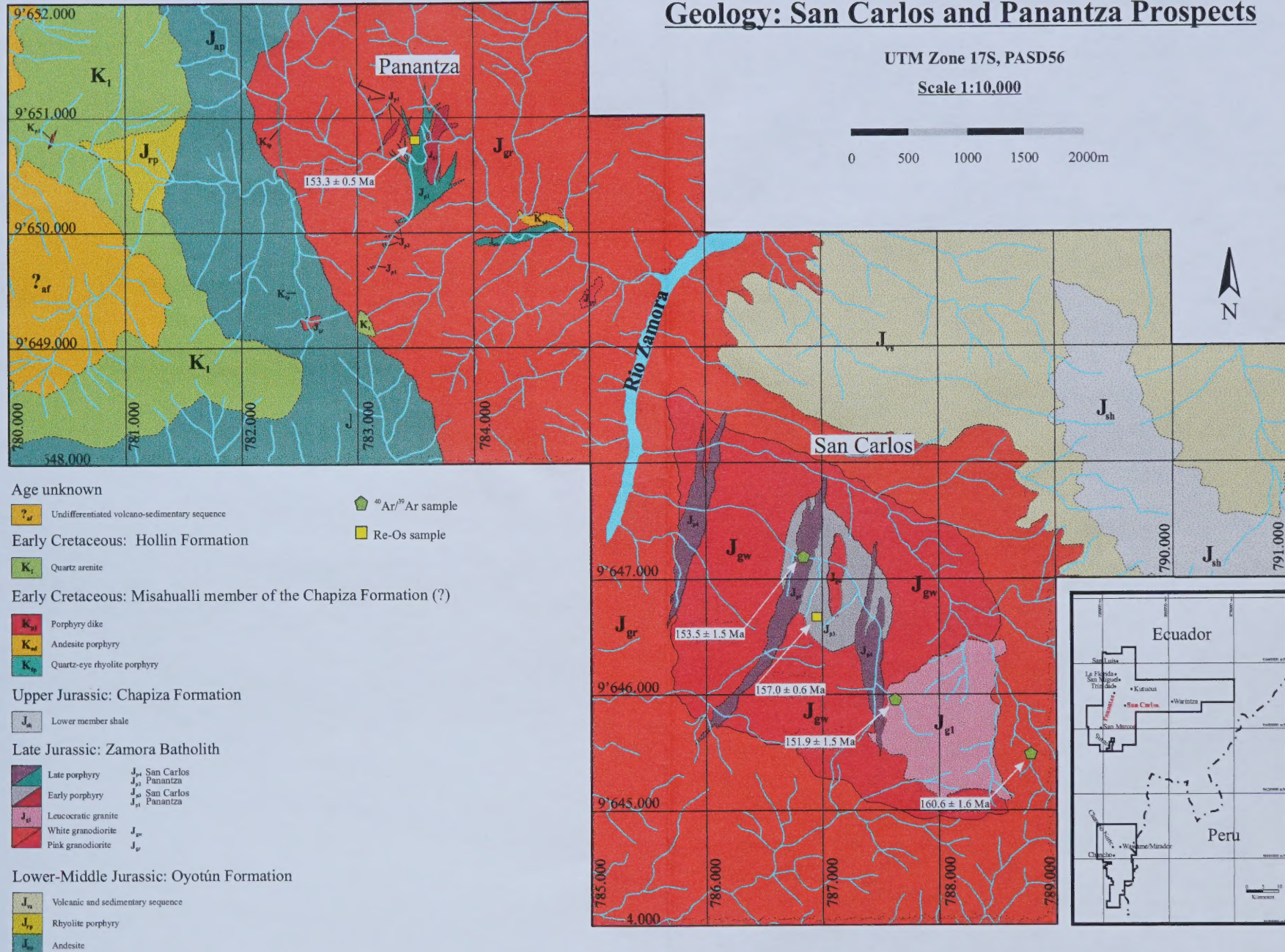
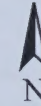
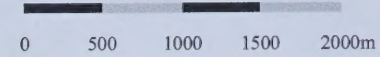
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